THE QUATERNARY STRATIGRAPHY AND SEDIMENTOLOGY OF THE KING RIVER VALLEY, WESTERN TASMANIA.

by

SEAN J. FITZSIMONS

BSc. (Hons) (University of Canterbury)

Submitted in fulfilment of the

requirements for the award of the degree of

Doctor of Philosophy

UNIVERSITY OF TASMANIA

HOBART

June 1988

Declaration.

This thesis contains no material which has been accepted for the award of any other degree or diploma in any university and contains no copy or paraphrase of material previously published or written by any other person, except where due reference is made in the text.

S J Fitzsimons

S. J. Fitzsimons.

ACKNOWLEDGEMENTS

I wish to thank several people for their assistance during this study:

- Prof. E. A. Colhoun for his criticism, advice and encouragement at all stages of the work.
- Dr. Guus van de Geer for his interest and work on the pollen from the organic deposits found in the King Valley.
- The Hydro Electric Commission of Tasmania which provided logistical and some financial support for fieldwork. In particular I thank Dr. F. J. Baynes and Mr. S. J. Paterson of the Geology section for their advice and access to drill core logs.
- Dr.R. S. Hill of the Botany Department for his work on the plant macrofossils.
- Dr. L. J. Wood of the Geography Department for proof reading some of the text.
- Mr. M. Pollington for permission to use his unpublished results on the palaeomagnetism of the glacial lake sediments of the King Valley.
- Dr. D. E. Lawson of the University of Waterloo for the use of his fabric analysis program.
- Dr. N. Eyles of the University of Toronto and Prof. G. S. Boulton of the University of Edinburgh for their advice with some problems in the interpretation of the patterns of sediment dispersal.
- The Australian Research Grants Committee for its support of this project.
- My colleagues in the Department of Geography, University of Tasmania for providing a friendly environment in which to work.
- Finally I thank Cathy for her support during the last three years.

ABSTRACT

This thesis is concerned with the Quaternary deposits of the King Valley in western Tasmania, Australia. The aims of the study were to describe the character and map the extent of the glacial deposits, to determine the stratigraphy and age of the deposits associated with each glacial advance, to compare the inferred processes and patterns of sediment genesis with the depositional processes observed at the margins of modern glaciers, and to formulate a model of environmental changes associated with the glaciations.

Deposits were mapped at a scale 1: 25,000 using morphostratigraphic, lithostratigraphic and biostratigraphic units. The stratigraphic classification and mapping defined thirteen formations from four glaciations. The last or Margaret Glaciation consisted of two main ice advances. The first appears to have occurred before 48ka; the second culminated after 19ka and had ended by 12ka. The Henty Glaciation, of middle Pleistocene age consisted of three ice advances. Two of the advances are separated by a long interval of lake sedimentation. The preceding Governor Glaciation, also of middle Pleistocene age, consisted of two ice advances. The glacial sediments are separated by organic sediments that record an interstadial climate. The Linda Glaciation is probably early Pleistocene in age. Its deposits are separated from younger glacial sediments by organic sediments that record the sucessional development of temperate rainforest during the Regency Interglacial. Linda Glaciation deposits are underlain by non-glacial sediments that are probably of late Pliocene age.

Detailed descriptions of the glacial sediments on which the stratigraphy was based have enabled elements of the dynamics and debris paths of the King Glacier to be reconstructed. These reconstructions suggest that the dominant sedimentary environment was supraglacial and that the debris was derived from the basal transport zone. Two of the more unusual patterns in the sediments are lithological stratification of erratic clasts in some tills and a widespread series of sedimentary wedges that resemble but differ from ice-wedge casts.

The chronology of the King Valley glaciations is estimated using a variety of relative dating methods that were applied to the deposits in their stratigraphic setting. The absence of reliable dating methods has meant that correlation of glacial events in Tasmania is largely based on comparing weathering characteristics. Wider correlation with glacial events in other Southern Hemisphere mid-latitude areas is not possible without accurate dating of the events.

CONTENTS.

LIST OF FIGURES

LIST OF TABLES

LIST OF APPENDICES

CHAPTER 1

GENERAL INTRODUCTION

1.1 Introduction

This thesis is concerned with the Quaternary glacial deposits of the King Valley in western Tasmania. The island of Tasmania is the southern-most state of Australia and lies between 40° and 44°S (Fig. 1.1). It experiences a temperate climate that is dominated by westerly maritime airstreams. Orographic uplift of the airstreams by the mountains of western Tasmania gives the area a high rainfall with annual averages as high as 3,600mm. The relatively low altitude of the mountains of Tasmania compared to those of New Zealand and South America means there are no glaciers or permanent snow. During the Quaternary however, Tasmania was glaciated on several occasions. One of the main centres of glaciation was the West Coast Range where the King Valley acted as a major outlet glacier for a small ice cap that formed on the Tyndall Plateau. The location of the valley leeward of the West Coast Range and the influence of topography on ice flow patterns have meant that the valley was a good area in which to expect a depositional record of the successive expansions of different glacial phases.

Early discoveries of the effect of ice action in Tasmania were made by a number of workers who recorded striations, erratics and moraines (Gould 1860, Sprent 1887, Dunn 1893, 1894, Johnston 1894, Moore 1894, Gregory 1904, Waller and David 1904, Ward 1909, David 1926a, 1926b).

The next major development in the study of glaciation was the work of A. N. Lewis from 1922 to 1945. Lewis used erosional evidence of former nivation levels to develop a multi-glacial model. He named the oldest and most extensive glacial episode the Malanna Glaciation and suggested that it was followed by two further glacial episodes of decreasing intensity. He called these episodes

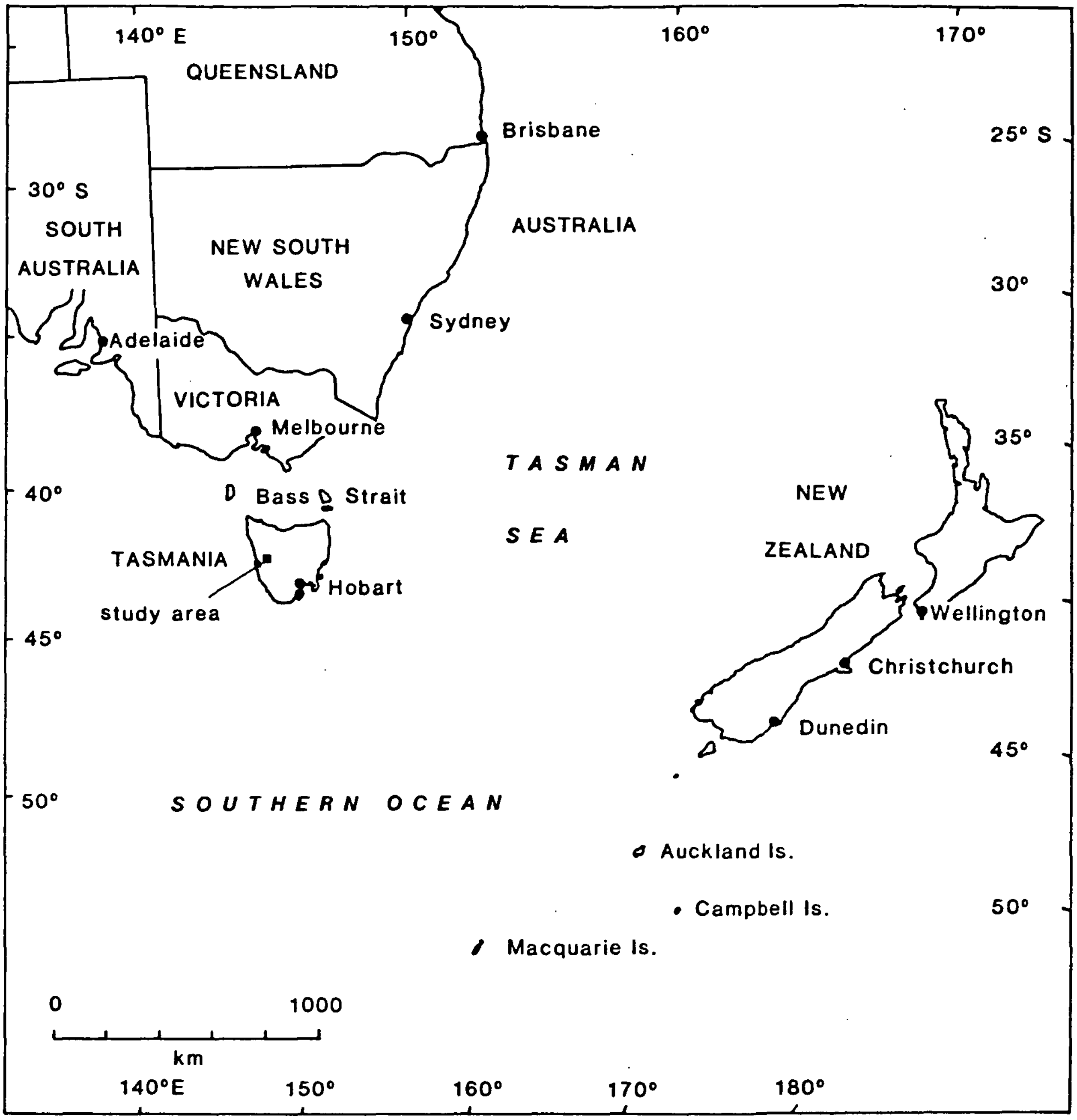


Fig. 1.1 Location map of Tasmania.

the Yolande and Margaret glaciations, and suggested that the three episodes correlated with the Mindel, Riss and Wurm Glaciations in Europe (Lewis 1922, 1926, 1934, 1939, 1945).

The three glaciation model developed by Lewis was succeeded by the work of Derbeyshire (1963), Davies (1962), and Peterson (1968, 1969) who studied the depositional evidence of glaciation. These investigations concluded that most of the glacial deposits could be attributed to one glaciation (Gill 1956, Jennings and Ahmad 1957, Jennings and Banks 1958, Banks and Ahmad 1959, Ahmad *et al.* 1959, Davies 1962, Peterson 1968, 1969). Derbyshire *et al.* (1965) summarised much of the evidence in the Glacial Map of Tasmania. Although the map was based on the recognition of only one glaciation, it conceded that erratics beyond the limits of continuous drift were possible evidence of earlier advances (Derbyshire *et al.* 1965, Derbyshire 1968, Derbyshire and Peterson 1971). In the Mersey-Forth area of northern Tasmania, Paterson (1965, 1966) and Paterson et al. (1967) recognised two glacial stages on the basis of the superposition of deposits, the degree of lithification and degree of chemical weathering. The older Lemonthyme Glaciation is the oldest glaciation known in Tasmania and is probably of Tertiary age.

The understanding of the glaciation of Tasmania changed little until the 1970's when ^{14}C dating and systematic mapping of the glacial deposits were combined. At Henty Bridge on the West Coast ^{14}C dating suggested that widespread glacial deposits predated the Last Glacial Maximum. (Banks *et al.* 1977). An increasingly complex stratigraphy emerged from subsequent studies (Bowden 1974, Colhoun 1976, Kiernan 1980, 1983a, Colhoun 1985b, Kiernan 1985). The location of the more recent stratigraphic studies is shown in Figure 1.2. By 1983 the accepted view of glaciation in Tasmania had returned to a three glaciation model. However, unlike Lewis' model it was based on the analysis of sedimentary evidence and supported by ^{14}C dating and pollen analysis of organic evidence.

This study involves the analysis of the glacial deposits in the upper King Valley, a relatively small area (150 km^2) in the central West Coast Range where abundant exposures were produced during dam construction work by the Hydro Electric Commission of Tasmania. This presented

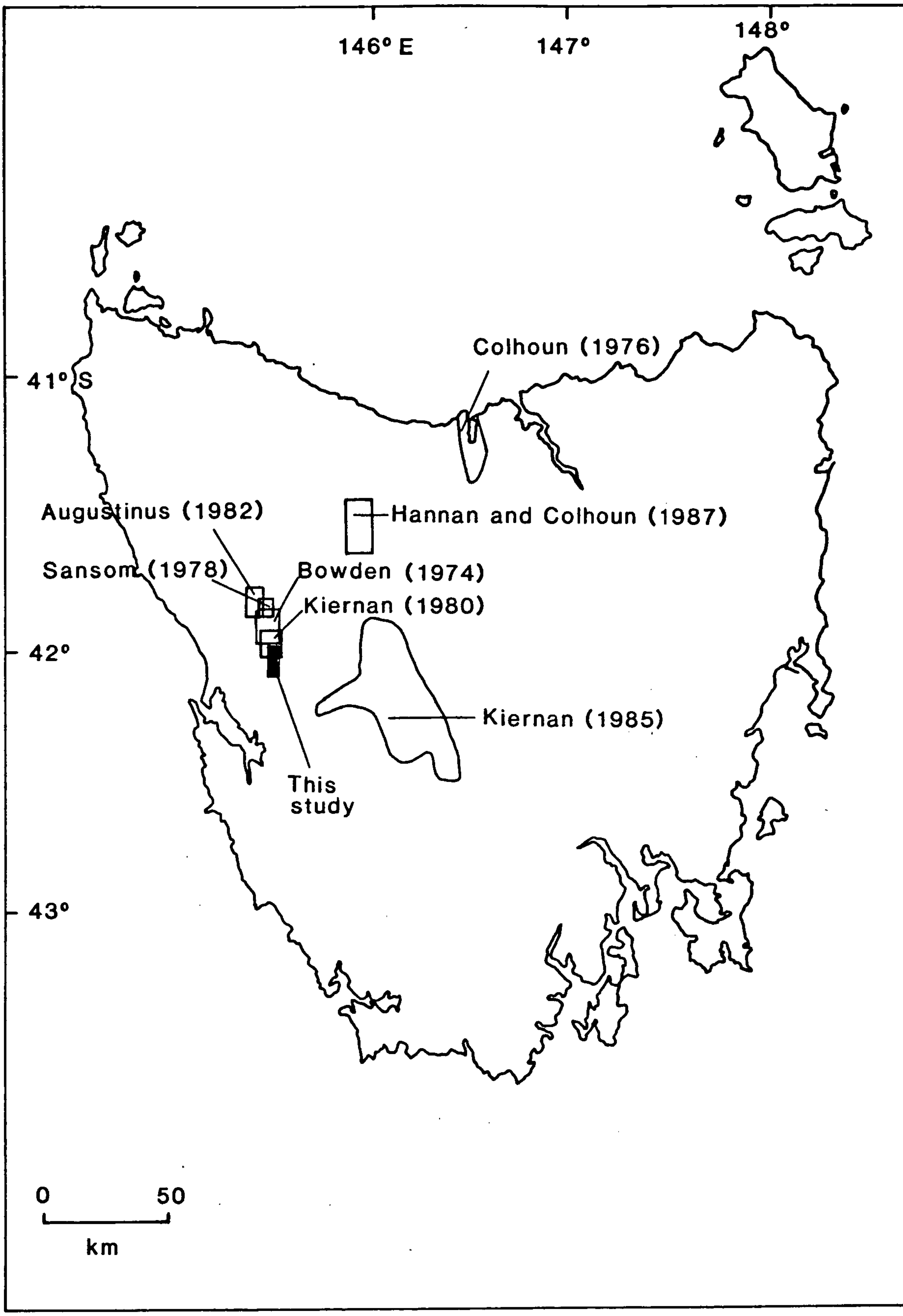


Fig. 1.2 Other Quaternary glacial sequences studied in Tasmania.

the opportunity to study the stratigraphy of the deposits near the terminus of a major outlet glacier system for the purposes of testing and possibly expanding the three glaciation model.

1.2 Aims

The aims and research questions addressed by this thesis fall into three groups and are summarised in Table 1.1.

The first group are aims that are concerned with establishing the stratigraphy of the glacial deposits of the King Valley, constructing a model of environmental change and comparing the stratigraphy with other areas.

The second group concerns the use of weathering rinds as a relative dating method. The use of this technique forms the basis of the existing three glaciation model for Tasmania and was developed by Kiernan (1980, 1983a). However, the question of the accuracy of using weathering rinds as a dating method in Tasmania has lacked serious testing. This study compares an independently developed stratigraphy with weathering rind data and discusses the strengths and limitations of the method.

The third group of questions have arisen from consideration of how modern glaciers can be used to interpret ancient deposits. The questions attempt to relate the inferred processes and patterns of sediment genesis to the glacial transport system and dynamics of the King Glacier.

1.3 Thesis structure.

This thesis is separated into three parts. Part One is an introduction and deals with the thesis aims, the physical environment of the King Valley and the methods that have been used. Part Two presents the stratigraphy of the Quaternary deposits of the King Valley, discusses the utility of weathering rinds as a dating method and describes the critical sections in the deposits. The first

TABLE 1.1 Summary of research questions and thesis aims.

1. AIMS RELATED TO THE FACTUAL DATA FROM THE STUDY AREA.
Establish the number and periods of glaciation in the King Valley (Chapter 4).
Determine the stratigraphy and extent of deposits associated with each glaciation (Chapter 4 and Map 1, back pocket).
Determine the character and extent of glacial deposits (Chapter 6 and Map 1, back pocket).
Using the morphostratigraphic and biological evidence, construct a model of environmental change associated with the periods of glaciation (Chapter 8).
Attempt a correlation of the glacial events known from the King Valley with events elsewhere in Tasmania and evaluate if they can be correlated with other southern hemisphere mid-latitude areas (Chapter 8).
2. QUESTIONS ARISING FROM THE USE OF PARTICULAR METHODS.
Is measurement of the thickness of weathering rinds on Jurassic dolerite an accurate relative dating method for the glacial deposits of the King Valley? (Chapter 5).
3. QUESTIONS ARISING FROM THE STUDY OF GLACIAL SEDIMENTS.
How are the inferred processes and patterns of sediment dispersal related to what is known about the patterns and processes of sediment genesis of modern glaciers? (Chapter 7).
Is subglacial lodgement till the dominant mode of deposition by temperate glaciers? (Chapter 7).

chapter of Part Three discusses the patterns and inferred processes of sediment genesis. In the second chapter a glacial chronology is presented and correlated with other glacial chronologies in Tasmania. An evaluation is also made of the prospects of correlation with the glacial events of other mid latitude Southern Hemisphere areas.

CHAPTER 2

THE STUDY AREA

2.1 Introduction

This study is primarily concerned with the upper part of the King Valley from the confluence of the Eldon and South Eldon rivers to the entrance of the King River Gorge through the West Coast Range (Fig. 2.1 and Map 2, back pocket). The area can be divided into five major physiographic units:

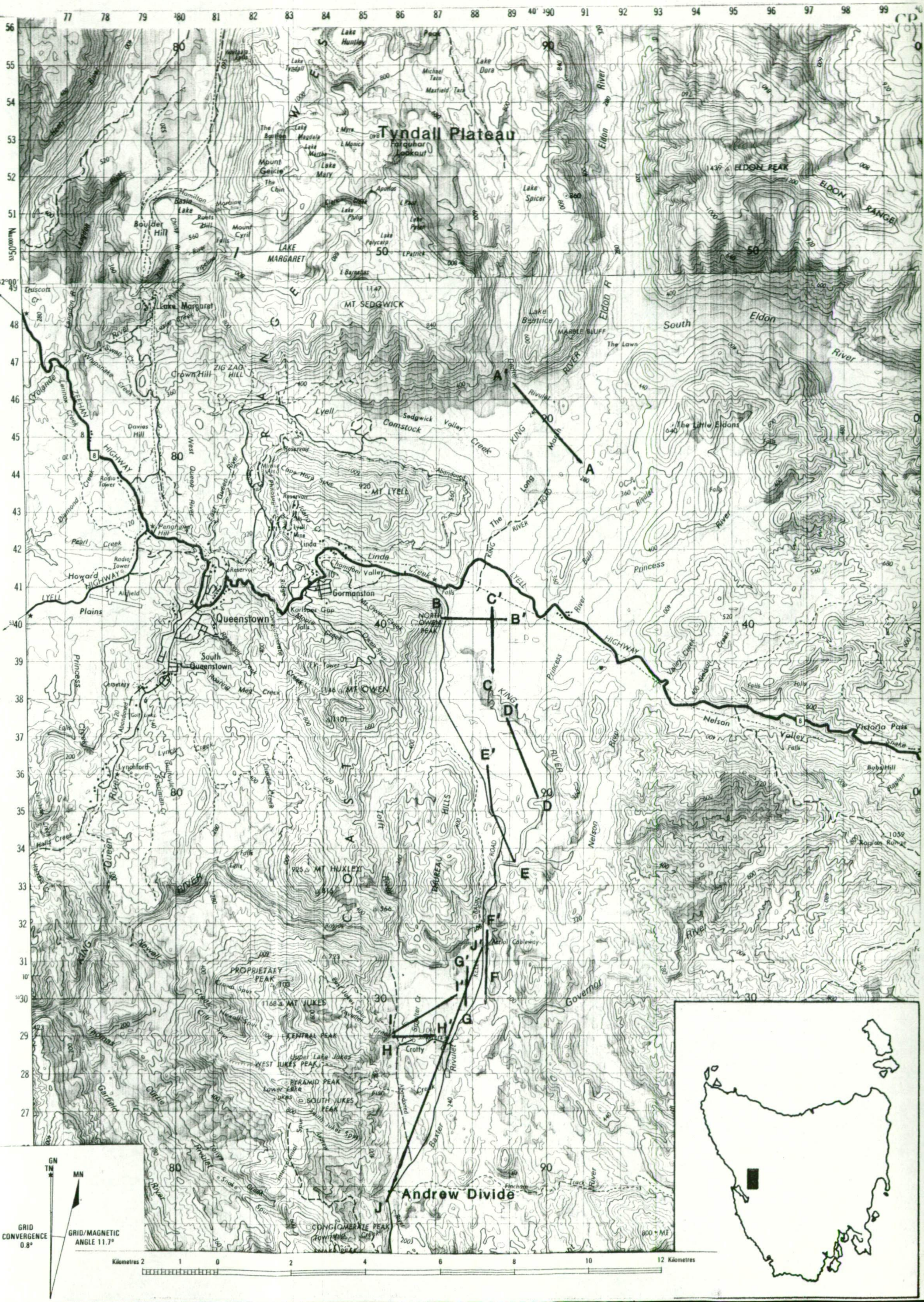
1. the Tyndall Plateau which forms a highland above 700m in the northern part of the study area;
2. the Eldon Range which forms an east-west trending mountain range up to 1439m high;
3. the West Coast Range which consists of a series of rugged north-south trending peaks range in altitude between 800 and 1275m (Fig. 2.1);
4. the King Valley which extends from the Eldon and South Eldon valleys down the eastern side of the West Coast Range;
5. the hill country to the east of the King Valley which forms an area of complex topography and is drained by the Governor and Nelson rivers.

The main physiographic features of the study area formed prior to glaciation and are, to a large extent, geologically controlled.

2.2 Geology

The King is a structurally controlled north-south trending valley in a geologically complex area (Fig. 2.2). It lies in a belt of complexly faulted and folded Cambrian, Ordovician and Siluro-Devonian rocks (Corbett *et al.* 1977). The northern part of the area is bounded by the Tyndall Plateau that consists of Precambrian schists, Permian sediments, Jurassic dolerite and

Fig. 2.1 Location map of the study area.



Tyndall Plateau
ELDON PEAK
ELDON RANGE
Lake Margaret
MT SEDGWICK
South Eldon River
Lyell Comstock Sedgwick Valley Creek
MT LYELL
Linda
Gormanston
Queenstown
South Queenstown
NORTH OWEN PEAK
MT OWEN
The Little Eldons
Princess River
Nelson Valley
Victoria Pass
LYELL HIGHWAY
MT HUXLEY
PROPRIETARY PEAK
MT JUKES
CENTRAL PEAK
WEST JUKES PEAK
PYRAMID PEAK
SOUTH JUKES PEAK
CONGLOMERATE PEAK
Crotty
Andrew Divide
KING RIVER
Governor
ZEEHAN
STRAHAN 32 km
GN
TN
MN
GRID CONVERGENCE 0.8°
GRID/MAGNETIC ANGLE 11.7°
Kilometres 2 1 0 2 4 6 8 10 12 Kilometres
A
A'
B
B'
C
C'
D
D'
E
E'
F
F'
G
G'
H
H'
I
I'
J
J'

Ordovician conglomerate. The eastern margin of the valley is bounded by faults that separate Siluro-Devonian sediments which crop out in the valley floor from Precambrian schists and quartzites. To the west, the valley is bounded by folded Cambrian and Ordovician rocks which form the West Coast Range (Fig. 2.1).

The Precambrian rocks consist of a wide range of meta-sediments that include quartzite, schists and phyllites. Although mostly outside the study area, clasts derived from Precambrian rocks form a substantial component of the Quaternary sediments.

The Cambrian Mt. Read Volcanics that form the core of the West Coast Range are a complex group of acid to intermediate volcanics (Corbett 1979). They consist mainly of tuffs, lavas, volcano-clastic conglomerate, greywacke and sandstone.

The Ordovician Owen Conglomerate forms most of the peaks of the West Coast Range. It overlies the Mt. Read Volcanics and forms an anticlinal structure. The sediments consist of a siliceous pebble conglomerate and sandstone with well cemented clasts composed almost entirely of Precambrian quartzite. The upper part of the Owen Conglomerate grades into the Gordon Limestone (Corbett *et al.* 1977).

The Ordovician Gordon Limestone occurs in low lying areas on both sides of the West Coast Range. It is composed of a medium to dark grey well indurated limestone with a calcium carbonate content of over 90%. In most areas it is weathered to a black clay which may be a residue of solution by acidic groundwater (Corbett *et al.* 1977). The limestone exhibits a karstic topography in a few areas including the Andrew Divide (F. J. Baynes pers. comm. 1987), the Nelson River and Dante Rivulet (Kiernan 1982, 1983b). Elsewhere, solution tunnels and dolines are buried and plugged by Quaternary sediments.

The Siluro-Devonian Eldon Group rocks conformably overlie the Gordon Limestone. They are composed of a wide variety of marine sediments that include quartz-sandstone, siltstone and shale. They crop out in the axial region of synclines such as the King Synclinorium (Gill

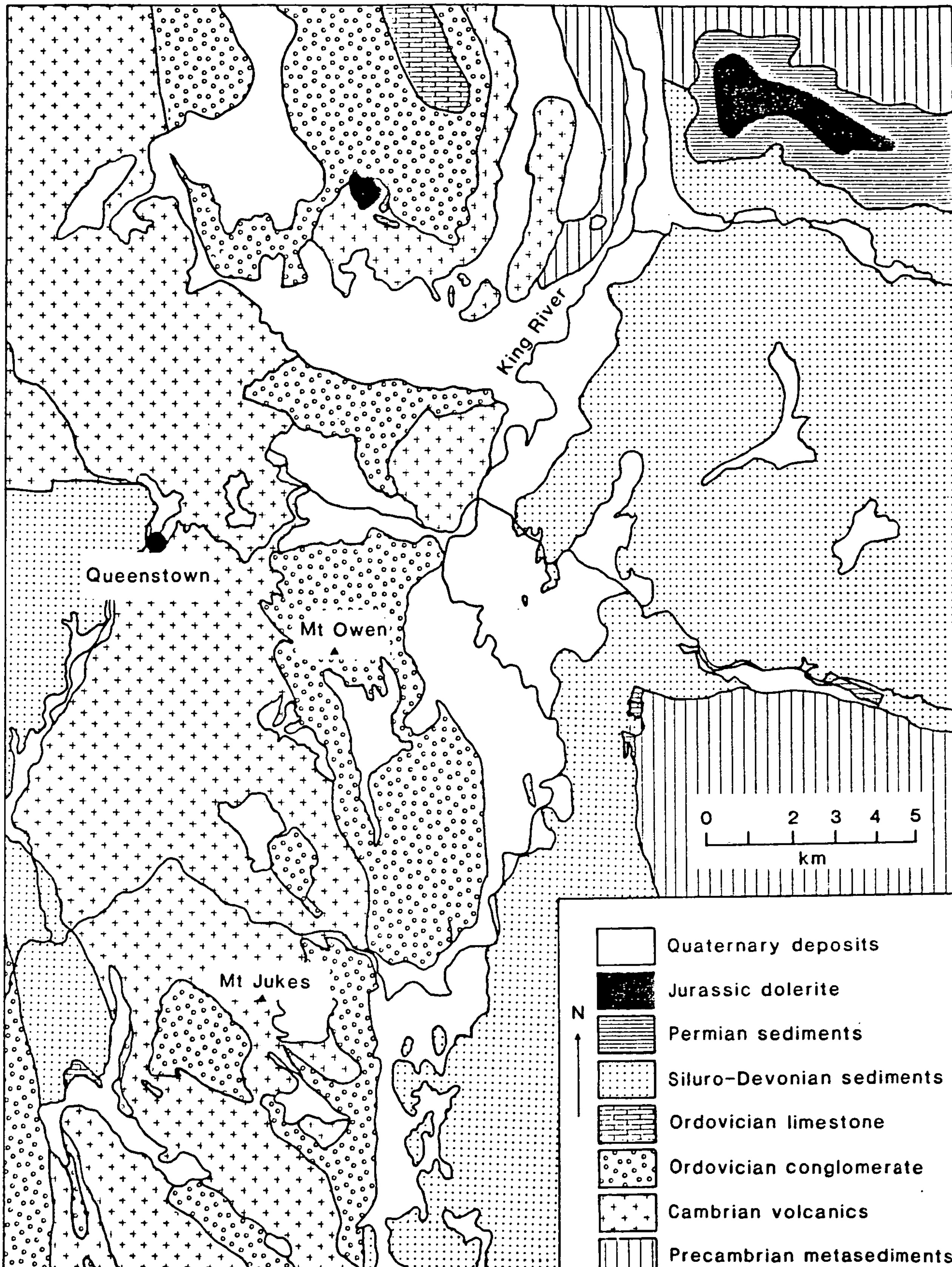


Fig. 2.2. Geological map of the study area. (from Corbett and Brown 1976 and Calver *et al.* 1987)

and Banks 1950). The Eldon Group contains six formations but by far the most common in the King Valley is the Crotty Quartzite which consists of a coarse quartz-sandstone that forms distinctive strike ridges in the central part of the valley.

Flat-lying Permo-Carboniferous sediments rest unconformably on a glaciated surface on folded Mt. Read Volcanics of the Eldon Range and Mt. Sedgwick. They consist of a basal tillite overlain by a sequence of conglomerate and calcareous siltstone (Banks and Ahmad 1962).

The Permo-Carboniferous sequence is intruded and baked by thick sills of Jurassic dolerite. The sills form distinctive caps on Mt Sedgwick and the Eldon Range.

Three distinct lithological provenances can be recognised in the King Valley. These are the dolerites and Permian sediments of the Eldon Range, the siliceous rocks of the West Coast Range and the local sediments derived from the valley floor (Table 2.1). The limited area of outcrop of both Permian sediments and Jurassic dolerite in the source areas for the King Glacier make them useful indicators of the direction of movement and extent attained by southward flowing glaciers in the valley. Similarly the absence of these rock types from sediments dominated by Owen Conglomerate and Mt Read volcanics can be used as evidence for derivation from the West Coast Range. The locally derived Eldon Group rocks constitute a significant component in almost all Quaternary sediments and therefore do not assist in defining the direction of ice movement or extent of glaciation.

2.3 Geomorphology.

The physiography of the study area is dominated by the West Coast Range which forms a north-south series of rugged peaks. The north-south continuity of the range is interrupted by the Comstock Valley, the Linda Valley and the King River Gorge (Fig. 2.1 and Map 2, back pocket). The heads of the Comstock and Linda valleys appear to have been breached and deepened by Quaternary glaciers. The King River Gorge appears not to have been occupied by glaciers. The northern part of the

TABLE 2.1 Lithology and source of rocks in Quaternary sediments in the King Valley.

ROCK TYPE	SOURCE
Jurassic Dolerite	Eldon Range (erratic)
Permian sediments	Eldon Range (erratic)
Siluro-Devonian sediments	Valley floor
Ordovician conglomerate	West Coast Range
Cambrian volcanics	West Coast Range

study area consists of the Eldon Range (1,200m) and the Tyndall Plateau (700m) which together with the northern part of the West Coast Range are the source area for the King Glacier.

The broad elements of the King Valley landscape are related mainly to variations in lithology and to fault and fold structures.

Most of the valleys are structurally controlled. The broad U-shaped, flat-bottomed King Valley is part of the King Synclinorium that is bounded to the west by the West Coast Range and to the east by the faulted contact with Precambrian rocks (Fig. 2.2). The central part of the King Synclinorium is occupied by the less resistant Siluro-Devonian Eldon Group sediments and the Gordon Limestone. The Linda and Comstock valleys are also structurally controlled by faults (Fig. 2.3). The courses of the Princess, Nelson and South Eldon rivers are controlled by the vertical dip, local variations in lithology and folded nature of the Eldon Group rocks.

A major feature of drainage that does not appear to be geologically controlled is the King River Gorge through the West Coast Range. This gorge is incised deeply into Ordovician conglomerate and Mt. Read Volcanic rocks, and crosses the major structural trends (Figs. 1.1 and 2.2). The gorge appears to be an antecedent valley which is characteristic of active orogenic areas (Bloom 1978). The study area is now known not to be orogenically active and the gorge is probably a relic form that may have developed during early Tertiary uplift of the West Coast Range.

The most impressive landforms of glacial erosion are the cirques and meltwater channels.

The main cirques formed on the eastern side of the West Coast Range and occur on Mts. Jukes, Owen and Geikie (Fig. 2.1). None of the cirques are deeply eroded and there is some suggestion that the altitudes of the cirque floors are geologically controlled (Bradley 1954). A few smaller cirques also formed on the Eldon Range. Although they face north and south the ones on the southern side of the range are considerably larger. The floors of these cirques appear to be at a similar altitude to the Late Palaeozoic-prePermian unconformity and may be also geologically controlled.

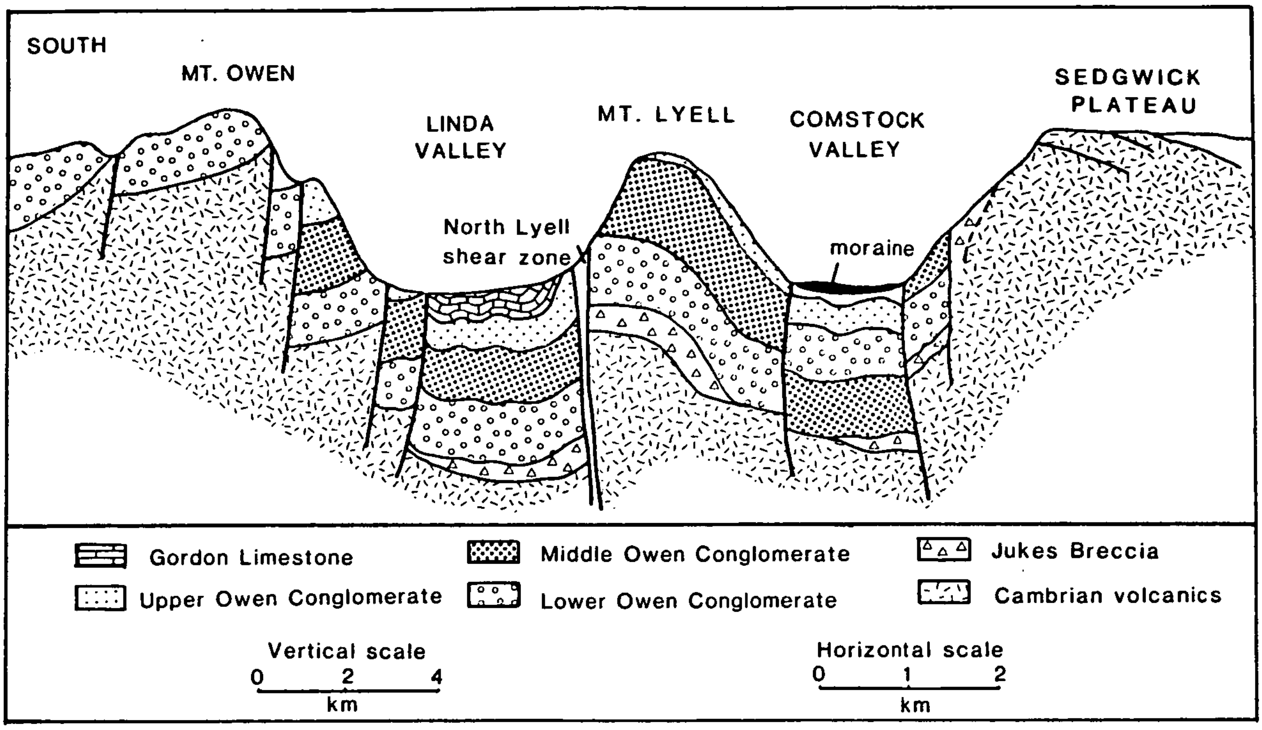


Fig. 2.3. Cross section of the Linda and Comstock valleys (from Solomon 1962).

Large meltwater channels cut in rock have been developed on Princess Ridge and on the eastern slopes of Mt. Lyell. Many of these channels have open intakes and cross ridges. They are likely to have formed when glacier ice rested against the ridges and forced meltwater streams to flow over the ridges and along the ice margin.

The major depositional landforms in the study area are outwash terraces and plains, screes and moraines. The central part of the valley is dominated by middle Pleistocene outwash plains (Fig. 2.4). The deposits of these plains are known to be up to 60m thick, and consist largely of outwash gravels and thick sequences of lake-bottom sediments. In the lower part of the valley, near the Governor River, outwash streams have deposited large amounts of gravel that have been reworked by successive outwash streams (Fig. 2.5). The resulting stratigraphy and terrace geometry is complex and records outwash gravels from numerous ice advances.

Most surviving moraines date from the Last Glaciation and are restricted to the higher cirques where they frequently dam small lakes (Fig. 1.1). On the floor of the valley there are two large moraine remnants. One is the Blackwood Formation end moraine which forms a low arcuate ridge across the valley (foreground of Fig. 2.4). The other is the Bull Formation end moraine which forms a ridge and outwash surface remnant about 3 km further north.

2.4 The glacial system.

The extents attained by glaciers during successive ice advances in the King Valley are shown on Figure 2.6.

Ice flowed into the King Valley from numerous cirques on the Eldon Range and West Coast Range, and from an ice cap on the Tyndall Plateau. During the more extensive advances the King Glacier split into four distributary lobes and flowed:

1. south down the King Valley;

Fig. 2.4. Outwash plain of the middle Pleistocene Blackwood Formation.

Fig. 2.5. Aggradation surfaces in the lower King Valley.

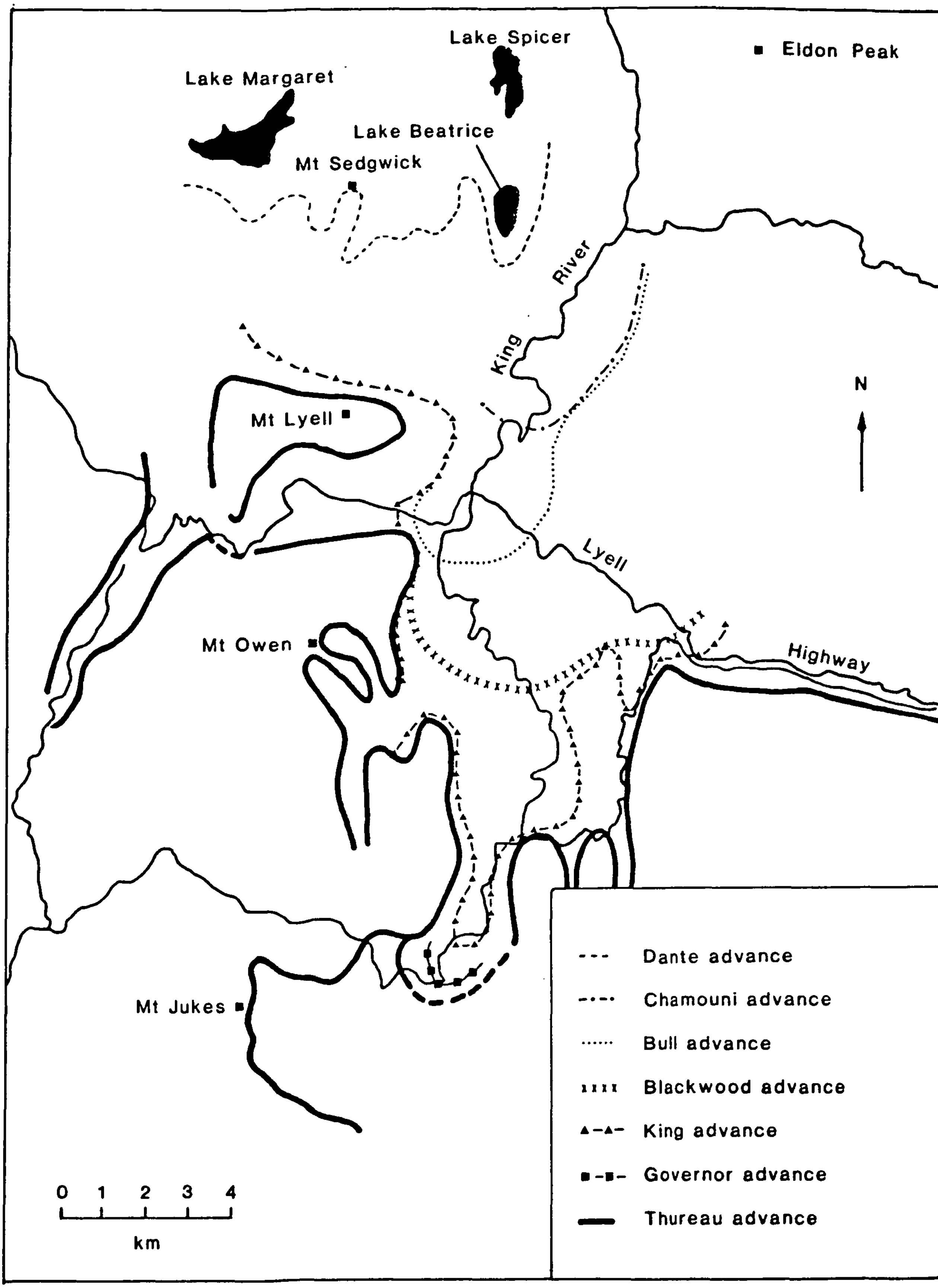


Fig. 2.6. Limits of Quaternary ice advances in the King Valley.

2. east and south into the Nelson Valley and into the Collingwood catchment;
3. west up the Comstock Valley;
4. west up the Linda Valley.

The Comstock and Linda distributary lobes appear to have coalesced west of the divide and flowed about 14 km down the Queen Valley (Kiernan 1980). The largest distributary lobe, which flowed down the King Valley, is the main concern of this thesis.

Although Kiernan (1980, 1985) suggested that ice from the King Glacier may have breached the Andrew Divide (Fig. 2.1) and flowed into the Franklin catchment, the absence of erratic Jurassic dolerite and Permian sediments in numerous exposures between the Governor River and the Andrew Divide, suggest this was not the case.

Several small glaciers formed on the West Coast Range and flowed down into the King Valley (Fig. 2.6). Ice from Mt. Owen flowed into the Tofft River where it coalesced with the King Glacier during its more extensive advances. Ice from Mt. Jukes, which carried the largest of the cirque glaciers on the southern part of the West Coast Range, flowed into the King Valley where it may have met ice from the King Glacier. The eastern extent of the Jukes Glacier is unknown but it seems likely that it extended into the Governor and Andrew river valleys.

The multiple ice sources that contributed to the glacial system in the King Valley makes the stratigraphic classification of the deposits complex. The local glaciers on Mt. Jukes and Mt. Owen were much smaller than the distant ice sheet which formed on the Tyndall Plateau - Eldon Range areas and fed the King Glacier. Consequently, the proximity of the local cirque glaciers to the lower King Valley meant that there was a time lag between the deposition of sediment fluxes associated with ice advances of the two glacial systems during the same phase of glacier advance. Andrews (1975) estimated the lag time for a small temperate glacier to be of the order of 10^2 years compared to 10^3 years for the Greenland ice sheet. The same principle applies to the glacial system of the King Valley though the sizes of the glaciers of the West Coast Range of Tasmania were much smaller. Although the volume of the glaciers cannot be accurately assessed, the King

Glacier is estimated to have been 25 larger in area than the Jukes Glacier. Because of the effects of time lag, the lithostratigraphic units from different sources but related to the same major climatic event can overlie or lie adjacent to each other. This phenomenon is recorded by a series of sections at Baxter Rivulet (section 6.5) where the Fish and Governor Formations appear to have been deposited during the same phase of glacial advance by ice from different sources.

2.5 Climate and Vegetation

Tasmania's position between approximately 40° and 44°S in the Southern Ocean and the Tasman Sea (Fig. 1.1) means that it experiences the coldest climate in Australia. The humid maritime climatic conditions are largely a consequence of the interception and orographic uplift of westerly maritime airstreams by the mountains. The West Coast Range area is the wettest region in temperate Australia with a mean annual rainfall of 2,000-3,600mm. The mean annual temperature at Queenstown is 10.9 °C. The mean monthly maximum is February at 22 °C and the mean monthly minimum is July at 2.3°C. There is no permanent snow or ice and the present permanent snowline is well above the highest peaks of the West Coast Range.

The natural vegetation of the King Valley would be temperate rainforest dominated by *Nothofagus cunninghamii* with a component of *Lagarostrobos franklinii* near rivers (Kirkpatrick 1977a). On the peaks of the West Coast Range the natural vegetation would be an alpine vegetation assemblage that includes the bolster plant *Donatia novae-zelandiae*, the coniferous shrubs, *Diselma archeri* and *Microcachrys tetragona*, the deciduous shrub *N. gunnii* and graminoids (Kirkpatrick and Dickinson 1984). However, because of the combined effects of poor drainage, extensive burning of the vegetation during exploration and selective logging activities in the area, the vegetation has been strongly altered (Kirkpatrick 1977b, Kirkpatrick and Dickinson 1984). Today the floor of the valley is mainly covered with eparidaceous heathland, *Gymnoschoenus sphaerocephalus* (buttongrass) sedgeland and wet scrub of *Leptospermum scoparium*, *Melaleuca squamea* and *M. squarrosa*. The better drained slopes have regenerating *Eucalyptus simondsii*. There is very little *N. cunninghamii* present except on river banks where it occurs with remnants of *L. franklinii* and is partially protected from burning.

CHAPTER 3

METHODS

3.1 Introduction

The methods used in this study can be divided into three groups, those employed in the mapping of the deposits and landforms, those used to differentiate the glacial advances and those used to reconstruct the environment of sedimentary deposition.

3.2 Mapping methods.

Although the study of glaciation in Tasmania has a long history (Banks *et al.* 1987), the stratigraphic classification and mapping of glacial sequences has attracted relatively little attention. Regional mapping by the Tasmanian Department of Mines has concentrated on defining pre-Tertiary rocks that have economic significance. Most mapping has been achieved through a number of honours theses (Bowden 1974; Sansom 1978; Kiernan 1980; Augustinus 1982). All of these lack a consistent approach to mapping and stratigraphic classification. This, and the difficulty of obtaining enough exposure in the forested and mountainous terrain have made both the identification and correlation of glacial formations, even over very short distances of the order of 10's of kilometres, difficult and in some cases speculative. Although none of the above studies resolved the complexity of the glacial stratigraphy, they have nevertheless provided the background against which the the present findings can be compared. The much greater quantity and detail of stratigraphic information available from the King Valley now permits a rigorous definition and classification of stratigraphic units.

The American Commission on Stratigraphic Nomenclature (1959) recommended the mapping of deposits using lithostratigraphic classification that can be related to periods of climatic change. The lithostratigraphic units are defined in the hierarchy of formation, member, lentil, tongue and bed, though in practice few studies of glacial deposits go beyond member because sedimentary complexity and rapid facies changes make small scale divisions of little meaning for purposes of wider correlation.

Although the ACSN recommendations regard morphologic units, "aggradational surfaces rather than bodies of rock", as invalid stratigraphic units (*ibid* p 665), many successful studies of glacial sequences have used morphologic units as primary mapping units. In New Zealand, moraines and outwash surfaces are frequently used as primary mapping units and their recognition may form the basis for defining a formation (Gage 1958, Suggate 1965a, Mabin 1984).

More recently, 'The International Stratigraphic Guide' (Hedberg 1976), has recognised the Quaternary as a special field of stratigraphy and suggests that lithostratigraphic classification based on indirect evidence such as geomorphology is a useful approach.

Moraines and outwash surfaces were the primary mapping units used in this study. However, because of the erosion of some primary depositional landforms it was necessary also to use geologically oriented methods. Several ice advances have been reconstructed from sedimentary evidence where there was an absence of geomorphic evidence. The mapping methods of this thesis adhere to the recommendations of the International Stratigraphic Guide where possible.

3.3 Criteria for differentiating glacial advances.

Several criteria have been used to separate glacial advances in the King Valley (Table 3.1). The simple criteria of altitude, and geographic distribution of moraines and outwash surfaces have been the most useful. However, because of the differential preservation of deposits, it has not been possible to use the same criteria to distinguish each formation.

Table 3.1 Criteria for differentiating glacial advances.

Advance	Distribution in King Valley	Preservation of surface form	Reconstruction of the sedimentary environment	Weathering (see Fig. 5.2)	Dating
Dante	Limited to the mouth of Comstock	Small outwash fan and moraine	Outwash gravel resting on older outwash gravel	Matrix unweathered	18,800± 550 yrs. BP (ANU 2533)
Chamouni	Isolated exposures in upper King Valley	Outwash terrace	Ice position reconstructed from location of till and outwash gravel	Slightly weathered	48,700 yrs. BP (SUA 2599)
Bull	Restricted to a small area 2km. south of Lyell highway	Moraine and outwash surface	No exposure	Moderately weathered	Not dated
Blackwood	Widespread south of the Lyell highway	Extensive arcuate moraine and large outwash surface	Advance beyond the position of end moraine inferred from sediments	Moderately weathered	Normal detrital remanent magnetisation Dated at 37,800 yrs. BP (SUA 2469.)
King	Restricted to south of the Governor River confluence	Small outwash terrace remnants and low hills of ice contact sediments	Environment and ice limit reconstructed from sediments.	Moderately weathered	Wood dated at 32,800 and 39,900 yrs. BP (SUA 2392 and 2393.) Regarded as minimum ages.
Governor	Restricted to the lower King Valley near the Governor River confluence	No surface form	Identified from the difference in weathering	Weathered	Not dated
Fish	Limited to the area south of the Governor River	Small remnants of terraces and ice contact ridges sedimentology	Distinguished from the Governor Formation by lithology and	Slightly weathered	Underlying Baxter Formation sediments have a normal detrital remanent magnetisation
Traveller	Limited to the area south of the Governor River	Single small terrace remnant	Distinguished from the Fish Creek Formation by the Baxter interstadial	Slightly weathered	Not dated
Thureau	Isolated remnants lower King Valley near the Governor River	No surface form	Separated from other sediments by the Regency interglacial	Highly weathered	Reversed detrital remanent magnetisation

Several dating methods including ^{14}C dating and palaeomagnetism have been used to supplement the geomorphic and sedimentary evidence. Data from weathering rinds on Jurassic dolerite were collected to test the utility of the technique as a dating method.

3.3.1 Relative altitude and distribution of aggradation surfaces.

Comparison of the relative altitude of aggradation surfaces is based on the assumption that the highest aggradation surface at a locality was deposited during the maximum extent of a glacier advance (Suggate 1965a). Identification of a series of outwash aggradation surfaces that can be traced upstream to either an abrupt steepening and termination of the surface or to a moraine, can be taken to reveal the occurrence of multiple ice advances.

Figure 4.2 shows the longitudinal profiles of outwash aggradation surfaces associated with the ice advances defined in Chapter 4. These profiles are drawn from 1:10,000 scale maps with contours plotted at either 5 m or 2 m intervals.

The geographic distribution of moraines can also assist interpretation of relative age because more extensive ice advances destroy landforms of earlier, less extensive advances. A statistical consideration of "obliterative overlap" by Gibbons *et al.* (1984) suggests that up to 7 out of 10 end moraines may be destroyed by a succession of ice advances. This interesting analysis serves as a warning that reliance on depositional morphology to determine the sequence of ice advances in a valley is likely to underestimate the number of advances. It is for this reason that reconstruction of depositional environments from fragmentary sedimentary evidence has proved to be an important tool in developing a glacial stratigraphy for the King Valley.

3.3.3 Palaeomagnetic dating.

Measurements of detrital remanent magnetisation of laminated lake sediments were made as part of a study of glacial lake sediments in Tasmania by Mr. Michael Pollington. The results reported in this thesis are preliminary and are quoted with permission as personal communications.

The importance of palaeomagnetism in defining the King Valley stratigraphy is threefold. Firstly, it has confirmed that some deposits beyond the range of ^{14}C dating are normally magnetised, and are therefore believed to have been deposited during the Brunhes Chron (0-730 ka. BP) (Bowen 1978). Secondly, deposits of the Thureau Formation are consistently magnetically reversed and are believed to have been deposited during the Matuyama Chron, ie older than 730 ka. BP. Thirdly, apparent changes in the palaeomagnetism recorded at site K28, the type section of the Thureau Formation (Fig. 6.4), appear to record sediments that cross a palaeomagnetic boundary, possibly from the late part of the Matuyama Chron into the Jaramillo normal event.

3.3.4 Radiocarbon dating.

The important ^{14}C dates in the King Valley and surrounding area are summarised in Table 3.2.

The value of ^{14}C dating in the development of the stratigraphy of the King Valley is limited because most glacial sediments are beyond the usual limits of the technique around 40 ka. However, ^{14}C dating has been useful in defining the Pleistocene-Holocene boundary and the limits of late last glaciation ice advance.

Although the stratigraphic position of many wood samples suggested they were well beyond the limits of ^{14}C dating, many were dated in an attempt to prove they really were beyond ^{14}C dating range. Most assays of this sort of material produced finite dates from 30 to 39 ka. B.P. (Table 3.2) even though the counter at The N.W.G. Macintosh Centre for Quaternary Dating at Sydney University has a nominal range of 50 ka. The problem is almost certainly due to contamination of the wood by modern humic acids which are mobile in humid environments such as the West Coast of Tasmania. This and other problems of contamination of datable materials in Tasmania have been outlined by Colhoun (1986).

Table 3.2 Radiocarbon dates in the study area

DATE (yrs BP)	LAB. CODE	LOCATION	SOURCE	COMMENTS
26,480 ± 800	W 323	Linda Creek G.R. 835422	Gill (1956)	wood from rhymites
>40,000	NZ 348	" "	Grant-Taylor &	
			Rafter (1962)	wood from proglacial
>40,000	R488	" "		Banks et. al. (1977)
>48,500	ANU 3413	" ".	Colhoun (1985b)	
27,800 ± 700	ANU 2480A	" "	"	
23,100 ± 600	ANU 2480B	" "	"	ß cellulose of ANU 2480A
9,050 ± 120	SUA 1358	Tyndall Plateau	Kiernan	
18,800 ± 500	ANU 2533	Dante Rivulet G.R.	Kiernan (1980)	Last glacial maximum
21,180 ± 370	SUA 2154	" "	Kiernan (1985)	
20,100 ± 470	SUA 2155	" "	"	
37,800 $^{+800}_{-700}$	SUA 2469	King Valley GR 872389	Colhoun &	
			van de Geer (1987)	
32,800 $^{+400}_{-700}$	SUA 2392	King Valley GR 870307	This thesis	Minimum age of Blackwood Formation outwash
39,300 $^{+800}_{-700}$	SUA 2393	" GR 875306	"	"
2150 ± 90	SUA 2487	Nelson River GR 917354	"	Recent river erosion
35,200 $^{+800}_{-700}$	SUA 2488	King Valley GR 878305	"	Minimum age of Blackwood
20,660 ± 280	SUA 2521	Newall Creek GR 800314	"	Peak of late last glaciation
2520 ± 80	SUA 2597	King Valley GR 881323	"	Holocene slope deposits
48,700 $^{+2900}_{-2100}$	SUA 2599	King Valley GR 886423	"	Minimum age of Chamouni Valley Formation
>21,000	SUA 2621	Newall Creek GR 800314	"	Basal date of organic sediments
12,250 ± 90	SUA 2415	King Valley GR 892378	"	Base of Holocene gravel
13,010 ± 130	SUA 2723	King Vallet GR 880307	"	End of the Last Glaciation

3.3.5 Weathering rinds on Jurassic dolerite.

Weathering rinds on Jurassic dolerite were measured to extend earlier work by Kiernan (1980, 1983a, 1983b) and to test the utility of the method on a relative stratigraphy determined by other methods. Fifty measurements were made at each site in accordance with the methods used by Colman and Pierce (1981) and Kiernan (1980, 1983a, 1985). Water absorption and specific gravity of clasts was also measured but these techniques were abandoned at an early stage because they provided the same information as the weathering rind data.
The results of this study are summarised in Chapter 5.

3.3.6 Pollen analysis of organic sediments.

The pollen analysis of organic sediments that were found in the study area have proved to be a valuable means of differentiating ice advances and glaciations. The detailed work on these sediments was done by Dr. G. van de Geer and Prof. E. A. Colhoun.

The pollen samples were prepared by the Faegri and Iversen (1975) method. To enable both percentage and concentration calculations to be made, exotic *Lycopodium clavatum* spores were added to the samples (Stockmarr 1971). A minimum pollen sum of 300 grains included all terrestrial woody and herbaceous taxa plus tree ferns. The taxonomic nomenclature follows Curtis (1963, 1967), Curtis and Morris (1975), Willis (1970) and Wakefield (1975).

3.4 Reconstructing the environment of deposition.

The large amount of exposure revealed during construction work in the King Valley provided a great deal of information for many sites. It was necessary to select that information that would be useful and critical in constructing a model for the glacial stratigraphy of the valley, in determining the sequence of events and interpreting the conditions of formation.

The data recorded from the sections exposed in the King Valley are partly quantative and partly qualitative. Qualitative aspects include mapping the geographic distribution of the deposits, studying the geometry of the deposits by recording two dimensional sections, and examining the lateral and vertical variations in the sequence of sediments, types of contact, primary and secondary sedimentary structures, the colour, weathering, and texture. Quantitative data, some directly comparable with studies of modern glacial sediments include; pebble fabric, particle size, weathering and stone count analyses.

Geometry.

The basic recording technique used in this thesis was to measure and draw two dimensional sections. All other recorded data were then plotted on the section diagram.

Structure.

Specific structures that include, faults, joints, folds, and a variety of post-depositional deformation structures were recorded on the measured sections wherever possible.

Colour.

Matrix colour described in this thesis was determined using "The standard soil colour chart" (Oyama and Takehara 1967). However, there is little to be gained in reporting the colour codes and the verbal codes are reported here.

Weathering.

In addition to weathering rinds a subjective classification of matrix weathering was used:

1. unweathered (Fresh). No visible matrix discolouration;
2. slightly weathered. Slight matrix discolouration, often associated with iron staining;
3. moderately weathered. Matrix discoloured, surface of igneous clasts decomposed;
4. highly weathered. Martix totally discoloured and iron stained, igneous clasts decomposed;
5. Completely weathered. Matrix and clasts completely decomposed, ghost rock structures.

Texture.

Several problems in the interpretation of particle size distributions make them an inappropriate method of environmental reconstruction (Selley 1970). The main problem in textural studies of the deposits of the West Coast of Tasmania is that most sediments predate the Last Glaciation and are intensely chemically weathered. The particle size of such sediments cannot be directly related to the process of transport or deposition.

However, particle size distributions were measured where a specific question was addressed such as the particle size of individual beds within laminated silt sequences. These samples were analysed by the Hydrometer method as summarised by Lewis (1984) and sieving.

Field determinations of particle size were made with a grain size comparitor and by measurement of larger particles with a tape. Sorting was assessed on the subjective scale of unsorted, very poorly sorted, poorly sorted, sorted, well sorted.

Pebble fabric.

Pebble fabric analysis of glacigenic sediments has been used to give two different types of information which include, direction of glacier movement, and the mode of sediment deposition (Boulton 1971). While the use of till fabric for the reconstruction of ice movement direction has a long history of use (West and Donner 1956), it is only recently that use of till fabric as a means of reconstructing the mode of deposition has become widespread.

The analysis of fabric data has suggested that particles become aligned parallel and transverse to glacier flow (Holmes 1941). Subsequently extensive studies of till fabric were used to infer regional ice flow directions (West and Donner, 1956).

The growing body of evidence that suggests that there is a strong relationship between mode of genesis and pebble fabric includes studies by Glen, Donner and West (1957), Mills (1977) Lawson (1979a), Haldorsen and Shaw (1982), and Doweswell and Sharp (1986). The transition from strong fabrics of melt-out and lodgement tills to the weaker fabric strengths of deformed and resedimented diamictons is interpreted as representing an increasing amount of disturbance or deformation subsequent to the release of sediments from ice (Lawson 1979a; Dowdeswell and Sharp 1986). The effects of deformation on englacial fabrics summarised by Doweswell and Sharp are:

1. a change from unimodal towards more random fabric patterns;
2. a reduction in S_1 (see below) and concurrent decrease in S_3 eigenvalues;
3. an increase in the frequency of steeply dipping clasts;
4. an increase in the variability in fabric orientation and an increased deviation of the principal eigenvector from the observed ice flow direction.

Because the local direction of ice flow in the King Valley is constrained by topography and known from numerous striae, it is possible to use pebble fabric to reconstruct the mode of sedimentation of the deposits.

Usually two or more sets of 25 samples were taken from an exposure the number of pebbles being based on Andrews and Smith's (1970) and Lawson's (1979a) sampling schemes that provide a maximum 5° variance at the 95% confidence level. Sample sets consist of prolate shaped pebbles which are defined as having axial ratios of $a/b \geq 2$ and $b \approx c$.

Measurements were plotted on equal area nets (lower hemisphere projection) and contoured using the method of Kamb (1959). The data were analysed using the eigenvector technique (Mark 1973). The largest eigenvector V_1 indicates the direction of maximum clustering and the mean axis. V_3 indicates the direction of minimum clustering. V_2 is normal to V_1 and V_3. The significance values, S_1, S_2, S_3, indicate the degree of clustering about the eigenvectors.

A cautionary note of many studies of pebble fabric is that while genetic interpretations based on fabric are possible, the technique is one of a number of methods described above that are best used in combination before a judgement on the mode of origin is made.

Stone counts.

The provenance of tills was determined by counting the lithology of at least 100 pebbles between 15 and 60 mm in diameter from each sample. The pebble counts were grouped into major lithostratigraphic units that crop out in the study area and are plotted on pie diagrams. The majority of unidentified clasts belong a group of Precambrian meta-sediments that have a wide variety of lithologies and are not well known.

Nomenclature.

There are numerous definitions of what actually constitutes a till (Dreimanis 1983, Dreimanis and Schluchter 1985, Lawson 1979a). The major issue of disagreement appears to be over the amount of reworking that can occur before the "glacial character" of the deposits is lost. The issue is avoided in this thesis by naming glacigenic sediments diamictons unless a specific interpretation is made.

CHAPTER 4.

STRATIGRAPHY.

4.1 Introduction.

This chapter summarises the Quaternary stratigraphy of the King Valley. Nine ice advances from four glaciations have been identified in the King Valley (Table 4.1). The extent and distribution of deposits associated with each advance is summarised in a map (Map 1, back pocket). The map describes and synthesises the nature and relationships between the multiple ice advances of several distributary ice lobes. The relationships between the deposits formed by the distributary lobes of the King Glacier and those formed by glaciers from Mts. Jukes and Owen are summarised by Table 4.2.

This temporal framework is used for the presentation of sedimentological data (Chapter 6), and its significance in the Tasmanian and wider context is discussed in Chapter 8.

The locations of the sections described in this thesis are given as grid references to Tasmap 1:100,000 Franklin sheet 8013 (Fig. 2.1).

4.2 The Gormanston Formation.

The type section of the Gormanston Formation is 1 km north of the town of Gormanston at G.R. 835422. The section shows a series of locally derived fluvial sands and gravels overlying a palaeosol with *in situ* tree stumps. Deposits of highly weathered Thureau Formation till occur on both sides of the section which is exposed on an eroded spur in a gully (Fig. 6.1). The local provenance and the dip of the beds suggest that the gravels are a locally derived foreset delta deposit. This deposit is eroded and is buried beneath Thureau Formation sediments, the oldest

Table 4.1 Formations of the King Valley and their interpretation

CLIMATIC STAGE	FORMATION	INTERPRETATION
Holocene	Long Marsh	post glacial
Margaret Glaciation	Dante Chamouni	glacial advance glacial advance
Henty Glaciation	Bull Rivulet Blackwood Nelson King	glacial advance glacial advance interstadial glacial advance
Governor Glaciation	Governor Fish Baxter Traveller	glacial advance glacial advance interstadial glacial advance
Regency Interglacial	Regency	interglacial
Linda Glaciation	Thureau	glacial advance

TABLE 4.2 Correlation of formations with deposits of the cirque glaciers and distributary lobes.

King Glacier	Mt. Jukes Glacier	Mt. Owen Glacier	Nelson River distributary	LindaValley distributary
LONG MARSH FORMATION	Minor stream bed instability	Minor slope deposits	Minor stream erosion	Minor slope deposits
DANTE FORMATION	Small outwash fans Small moraines in cirques Moraine dammed lakes	Small outwash fans Terminal moraine in cirque	Erosion of older sediments	Slope deposits and minor erosion
CHAMOUNI FORMATION				
BULL FORMATION				
BLACKWOOD FORMATION				
NELSON FORMATION	Not found	Not found		
KING FORMATION			Numerous end moraines and extensive glacial deposits	Not differentiated
GOVERNOR FORMATION	FISH FORMATION			
Not found	BAXTER INTERSTADIAL		Not differentiated	
Not found	TRAVELLER FORMATION			
REGENCY FORMATION	Not found	Not found	Not found	Organic remains at Lynchford
THUREAU FORMATION	Extensive terraces in Governor and Andrew Rivers	Not found	Extensive, highly weathered ice-contact deposits	Large moraines in the head of the Linda and Queen Valleys
Erosion surface	Not found	Not found	Not found	GORMANSTON FORMATION

known glacial deposits in the study area.

Initial dating of wood from the palaeosol gave a date of 26,480 ± 800 yrs. B.P. (W-323), (Gill 1956). For many years this date was considered to record the maximum extent of ice during the last glaciation. Two subsequent dates of >40,000 yrs. B.P. (R-488 and NZ-348), (M Banks pers. comm., Grant-Taylor and Rafter 1963), contradict the earlier date and suggest problems in interpretation. Recent dating of whole wood at 27,800 ± 700 yrs. B.P. and the alpha cellulose fraction at 23,100 ± 600 yrs. B.P. suggested a serious contamination problem (Colhoun 1985a). Determination of the $o^{13}C/o^{12}C$ values of the dated wood suggests that all ^{14}C dates are suspect (*ibid* p. 54). Although the dating problem is not completely resolved it appears that the wood is probably of infinite radiocarbon age.

Earlier study of pollen from the palaeosol showed that the deposit contained Tertiary floral relicts including the extinct *Nothofagus brassii* type pollen taxon. (Kiernan 1980), but it was interpreted as a derived pollen assemblage.

Although it was previously thought that the sediments were of glacial origin (Colhoun 1985a), study of the lithology, geometry and sedimentology of the section, described in Chapter 6, suggests that the whole section is probably of Tertiary age.

Sediments underlying the Thureau Formation are not known elsewhere in the study area.

4.3 The Thureau Formation.

Thureau Formation sediments are preserved in five areas, the Linda Valley, the Queen Valley, the lower King Valley, the Nelson Valley, and the middle King Valley near the Thureau Hills (Map 1, back pocket and Table 4.2). Most of these occurrences are beyond the limits of subsequent middle Pleistocene ice advances.

The type section of the Thureau Formation is exposed at G.R. 882353, on the eastern slopes of the Thureau Hills. The section consists of glacial lake sediments and multiple sediment flows (Fig. 6.4). All deposits of this formation in the King Valley are buried beneath slope deposits or outwash gravels of subsequent ice advances.

During the Thureau advance ice reached its maximum extent in the King Valley and split into four distributary lobes. (Fig. 2.6). At this time ice extended 19 km down the King Valley to the entrance of the King River Gorge through the West Coast Range where it terminated near or against ice from Mt. Jukes. The King River Gorge acted as the major meltwater channel and carried large amounts of sand and gravel that accumulated on the western side of the West Coast Range near Newall Creek.

Reversed detrital remanent magnetisation of lake silts of the Thureau Formation at Thureau Hills suggest that the formation predates the Brunhes-Matuyama boundary and is older than 730,000 yrs. B.P. (M. Pollington pers. comm. 1986)

In the lower King Valley sediments of the Thureau Formation are considerably more weathered than those further up the valley near the type section. Although the different degrees of weathering might be taken to suggest that a substantial period of time separated their deposition, the evidence is insufficient to classify the more weathered sediments as belonging to a different advance. They are therefore mapped as one formation.

4.4 The Regency Formation.

The Regency Formation overlies Thureau Formation glacial deposits and consists of an organic deposit that contains an interglacial flora.

The type section of the Regency Formation is exposed in a gravel quarry near the confluence of the King and Governor Rivers at G.R. 875311. The section consists of 1.1 m of organic sediments that lie above highly weathered Thureau Formation till and below 5 m of Blackwood Formation outwash gravel.

Pollen analysis and examination of plant macrofossils of the Regency Formation shows an interglacial flora rich in *Lagarostrobos franklinii, Nothofagus cunninghamii and Phyllocladus aspleniifolius* (G. van de Geer and R. Hill pers. comm., Fig.4.1).

The relative pollen diagram (Fig. 4.1) can be divided into two zones RY1 and RY2 between 35 and 40 cm. The main difference between the zones is the very high percentages of pollen of temperate rainforest trees, particularly *Lagarostrobos franklinii* above 35 cm. and the greater sclerophyll component indicated by *Eucalyptus* and *Casuarina* below 35 cm. In addition, above 35 cm *Nothofagus* peaks, *Eucryphia-Anodopetalum* and *Anopterus* occur, there are abundant spore of ground ferns and marked peaks of the tree ferns *Cythea* and *Dicksonia*. These taxa are largely absent below 35 cm where high values of *Pomaderris*, *Arthrotaxis*, *Microstrobos* and *Coprosma* occur.

In western Tasmania substantial quantities of *Pomaderris* usually occur as understorey to wet sclerophyll *Eucalyptus* or in mixed *Eucalyptus* rainforest. *Arthrotaxis* is best represented in montane rainforest. *Microstrobos* is a subalpine shrub and *Coprosma* is usually associated with subalpine plants in Tasmanian pollen diagrams.

Although the boundary between RY1 and RY2 is drawn between 35 and 40 cm it is not a sharply defined biostratigraphic boundary; rather it is a transition. The transitional nature is indicated by the reduction of the higher altitude components *Arthrotaxis* and *Microstrobos* at 50 cm while *Eucalyptus* and *Casuarina* do not peak until 40 cm. It is difficult to judge what continuity of pattern in the vegetation changes may have been lost by the absence of pollen from 45 cm. (E. A. Colhoun pers. comm. 1987).

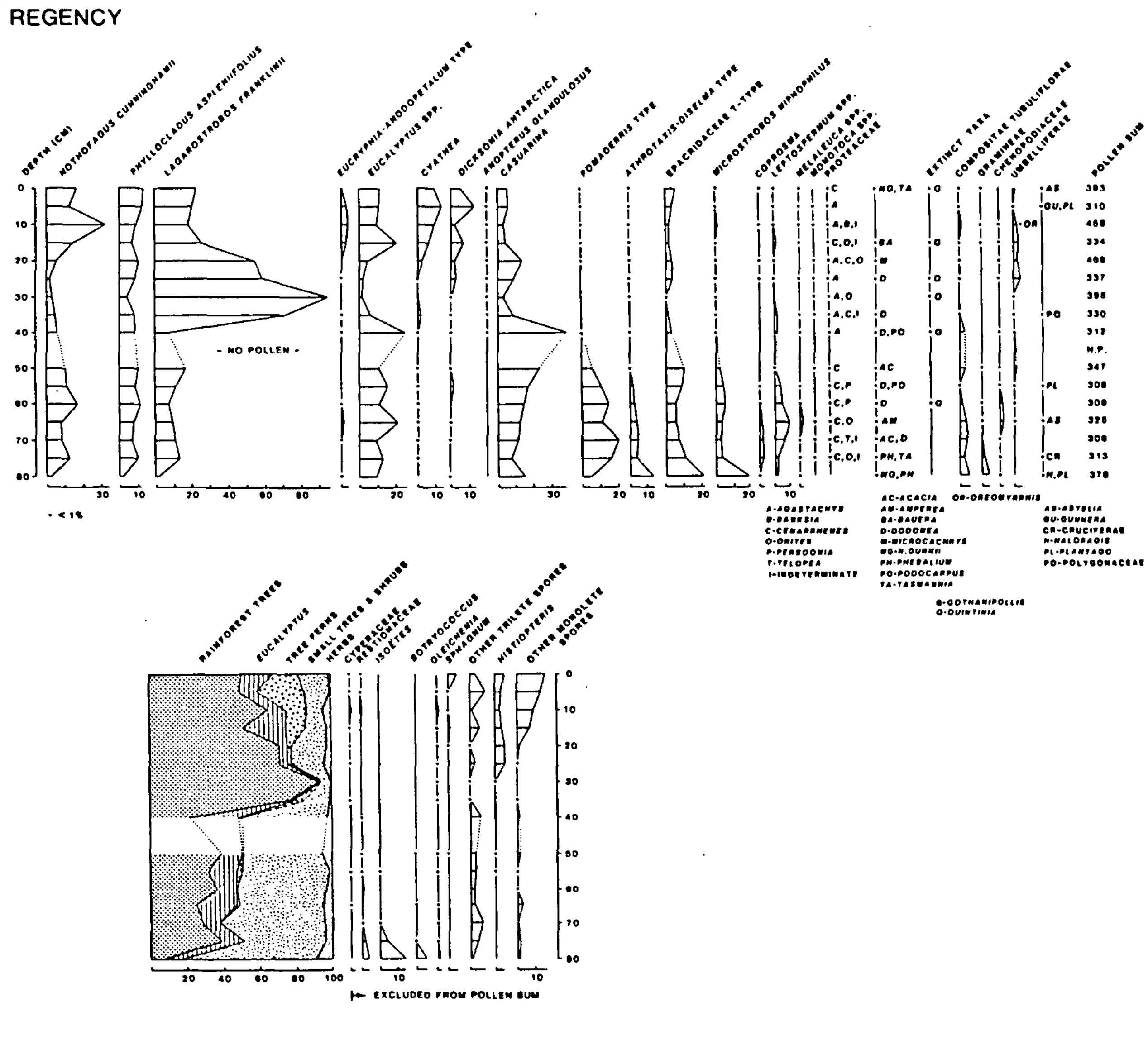


Fig. 4.1. Pollen diagram of the organic sediments of the Regency Formation.

Two of the most interesting taxa recorded are the extinct species (for Tasmania) *Quintinia psilatispora* and *Gothanipollis perplexus*. Today *Quintinia* is a subtropical wet forest taxon and *G. perplexus* is a wet forest parasite. *Quintinia* has not previously been recorded from Tasmania in deposits of Pleistocene age. The youngest previousl known record is from the fossil soil buried in the Gormanston Formation that is believed to be older than the Thureau Formation which is over 730,000 years old. The Gormanston record is either of early Pleistocene or late Tertiary age, but the present record is stratigraphically younger than the Thureau Formation. The record suggests that *Quintinia* may have survived in Tasmania until after glaciation had commenced.

The interpretation of the pollen data and the vegetation assemblages that have been inferred indicate that the climate was wet at all times. The large sclerophyll and small subalpine components of RY2 suggest that the vegetation had not yet developed to the optimum condition for the region of temperate lowland rainforest, but that it was either successional or adjusted to cooler montane conditions, or both.

Pollen zone RY1 contrasts markedly with RY2 with the expansion of *Lagarostrobos* to very high values. Treeferns and ferns also became important, and the sclerophyll, heath and subalpine taxa decrease markedly. The high values for *Lagarostrobos*, *Nothofagus* and *Phyllocladus* indicate that RY1 represents a lowland temperate rainforest with a marked riparian component. This pollen zone indicates that the organic deposit between the two glacial deposits is not only stratigraphically, but biostratigraphically interglacial, and that it represents part of an interglacial optimum. To what extent the decrease in *Lagarostrobos* in the upper 15 cm reflects the commencement of deterioration of climate and/or soil during the later part of an interglacial cycle, as possibly suggested by the advent of *Eucryphia*, is difficult to say as the sequence has been truncated by meltwater erosion. However, the lower part of RY1 must climatically represent a lowland temperate rainforest environment that was a least as wet and warm as the maximum of the Holocene.

4.5 The Traveller Formation.

The Traveller Formation was deposited by an ice advance of the Jukes Glacier. The type section is exposed on the right bank of Baxter Rivulet at G.R. 875300 and consists of 1.5 m of coarse outwash gravel resting on remnants of the Thureau Formation and on weathered Ordovician limestone.

This formation has a very limited extent as it is overlain by the more recent Fish Formation. It is not possible to differentiate between the Traveller and Fish Formations in outcrop where they are not separated by the Baxter Formation because they are lithologically and sedimentologically identical. Profile G - G', (Fig. 4.2) shows a high terrace which is the only remaining surface expression of the Traveller Formation. The terrace consists of 6 m of bouldery outwash gravel that lies well above the aggradation level of the Fish Formation.

Sediments of the Traveller Formation are probably a great deal more extensive than mapped because it is very difficult to distinguish them from the Fish Formation. Where this confusion occurs, the deposits have been mapped as an undifferentiated part of the Governor Glaciation (Map 1).

4.6 The Baxter Formation.

The Baxter Formation consists of 1.2 m of organic silty sand. Pollen analysis of the silty sand shows an herbaceous assemblage with an alpine component. The type section is the same as for the Traveller and Fish formations.

The relative pollen diagram (Fig. 4.3) can be divided into two zones, BR1 and BR2, above and below 40 cm respectively. The zone boundary is not well defined by any important taxon or group of taxa, but is placed having regard to overall changes in the pollen assemblages, and by inference the composition and structure of the vegetation.

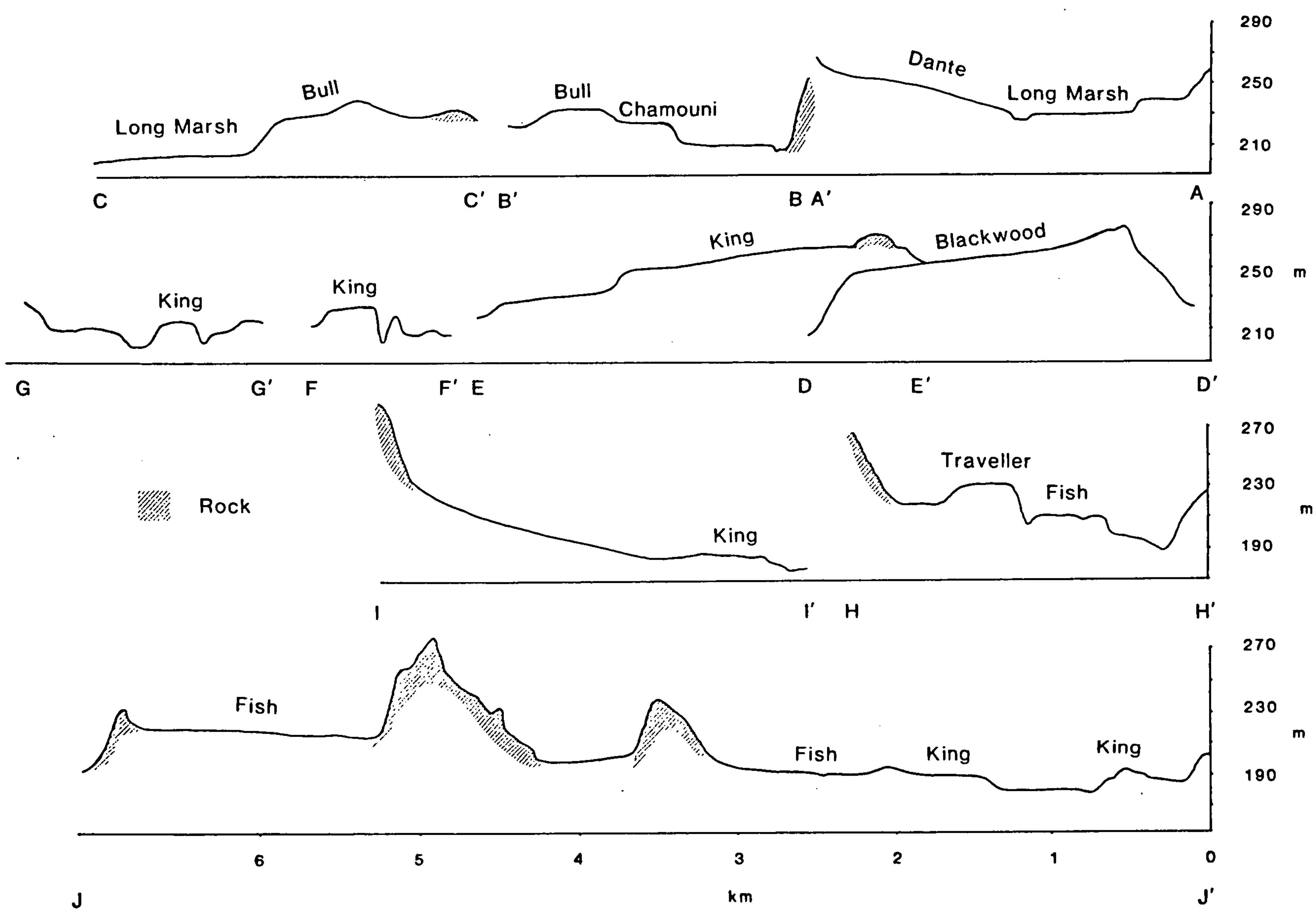


Fig. 4.2. Long profiles of aggradation surfaces in the King Valley. The profiles are drawn from 1:10,000 scale maps with contours plotted at 5 m or 2 m intervals. Locations of the profiles are shown by Fig. 2.1.

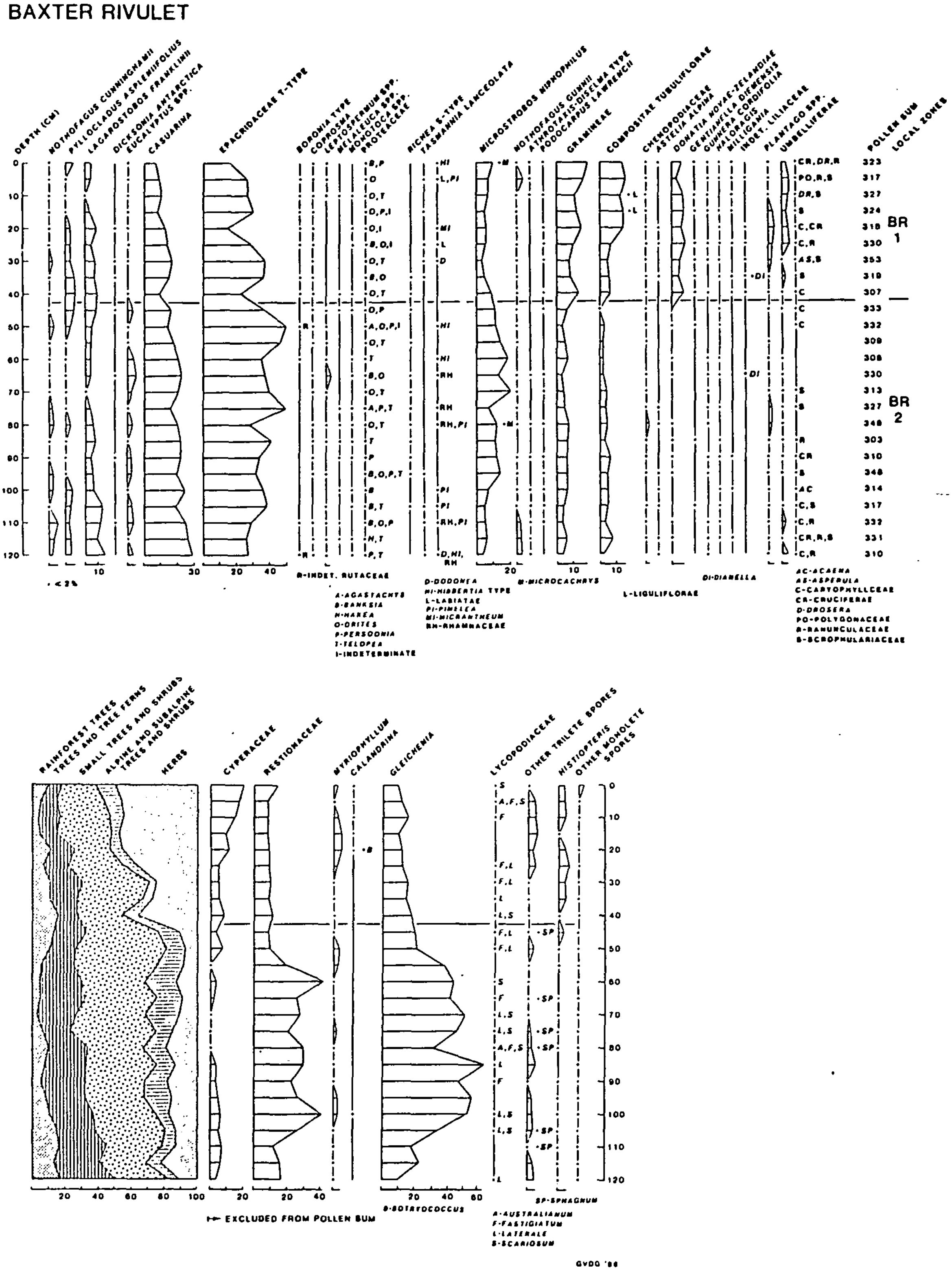


Fig. 4.3. Pollen diagram from the Baxter Formation.

Pollen zone BR1 (0-40 cm) is characterised by the maximum development of Gramineae (19%) and Compositae (15%), with 3-8% *Donatia novae-zelandiae, Plantago* and Umbelliferae. In this zone the percentage for herbs is always greater than the mean for the profile.

Pollen zone BR2 is characterised by high values of Epacridaceae (50% at 50 cm), *Casuarina* (29% at 120 cm) and *Microstrobos* (20% at 70 cm). In contrast to zone BR1, tree and alpine and subalpine shrub pollen values are greater than the mean values of 20% and 11% for the profile respectively, except below 100 cm where alpine and subalpine pollen decrease below 9%.

The two major temperate rainforest taxa *Nothofagus cunninghamii* and *Phyllocladus aspleniifolius* never exceed 8% collectively, and generally do not exceed 4%. If temperate rainforest was present these values would almost certainly exceed 25%. The values for *Lagarostrobos franklinii* are also generally low and only exceed 10% at the base of the profile. These low values suggest the presence of occasional trees or small stands of trees close to the river bank. The very low values of *Dicksonia antarctica*, which do not exceed 0.3% on any horizon, also suggest that all taxa usually found in or adjacent to rainforest were near their upper limits of distribution or were absent from the area. Although *Eucalyptus* pollen is more likely to be derived locally than that of rainforest taxa, the low values not exceeding 5% suggest there were few eucalypts.

Most of the pollen is derived from Epacridaceae (T-type), *Casuarina* cf *monilifera* and *Microstrobos niphophilus* with small quantities from Gramineae and Compositae. The most striking pattern is the gradual decrease in *Casuarina* from 120 to 0 cm, and the slight relative increases of Epacridaceae and *Microstrobos* in the upper part of zone BR2. These three pollen taxa contribute mean values of 70% in zone BR2 and only 46% in zone BR1. Though both *Casuarina* and *Microstrobos* are capable of long distance transport in small quantities, the values recorded here indicate that these taxa formed important components of the vegetation.

Towards the top of the profile in zone BR1, the marked increase in Compositae and Gramineae pollen are accompanied by *Donatia novae-zelandiae* (8% at 30 cm) in association with *Plantago* and Umbelliferae. *Donatia* pollen rarely occurs in quantity unless the plant is growing in the immediate locality which is normally at altitudes of over 1000 m today, though in western Tasmania cushions occur at lower altitude if free from competition from other plants. However, the quantity of *Donatia* and the association of the five pollen groups suggest they formed part of a higher altitude assemblage than could not occur at the site today.

The high values for Restionaceae and Cyperaceae indicate a locally wet habitat on the accumulating flood plain. Such an interpretation is strongly supported by the very high values of *Gleichenia* which grows abundantly on low river banks in western Tasmania. In addition, the almost consistent presence of *Myriophyllum* shows the occurrence of standing water which probably formed swampy pools on the flood plain.

Pollen concentrations vary from a minimum of 3.3×10^3 gr/g (grains per gram) at 10 cm to a maximum of 30.7×10^3 gr/g at 45 cm, but there is no overall difference between zones BR1 (8.5×10^3 gr/g) and BR2 (8.7×10^3 gr/g). The concentration of pollen appears to be related more to the organic content of the fluvial sediments than to systematic changes that can be interpreted in terms of pollen production/vegetation density.

Charcoal fragments >20 μ in size occur in all samples and show highly variable values which is characteristic of fluvial sediments. The low values of charcoal throughout the profile indicate that fire is unlikely to have been an important factor in either establishing or altering the inferred structure of the vegetation communities, but occurred in the catchment throughout the period of deposition of the sediments.

The pollen evidence indicates that during zone BR2 the vegetation was either an open *Casuarina* woodland with an abundance of *Microstrobos* and Epacridaceae shrubs or a wet shrub-heathland in which *Casuarina* was abundant. It was neither a forested nor a predominantly herbaceous vegetation. The vegetation became more open during zone BR1 as the average proportion of woody taxa was reduced from 86.4% to 60% and herbaceous taxa, particularly Gramineae and Compositae, increased from 13.6% to 40%. The change suggests that either the understory of the woodland became grassy or that a mosaic of scrubland-herbland-heathland vegetation was developed.

The preferred interpretation is that the vegetation was a wet *Casuarina* heath which either became a herbland-heath mosaic or a mix of herb and heath species (E. A. Colhoun pers. comm. 1987). This inferred vegetation differs from other non-forest pollen vegetation assemblages recorded from lowland western Tasmania in the abundance of *Casuarina*.

The flora is non glacial and lies between two glacial outwash gravels. It records an interstadial event *sensu stricto*, that is a non glacial period in a glacial event (Bowen 1978). The interstadial is known only from this site.

Dates on wood in the overlying King Formation outwash gravel are considered to be infinite because they are close to ^{14}C background (Table 3.2). No attempt was therefore made to carbon date wood found in the Baxter Formation. The Baxter Formation is certainly beyond ^{14}C age and is probably of middle Pleistocene age.

4.7 Fish Formation.

The Fish Formation was deposited by an ice advance of the Jukes Glacier, and is named after Fish Creek where large sections of ice contact sediments are exposed.

The type section is the same as that of the Baxter Formation. At this section the Fish Formation is represented by 4 m of coarse outwash gravel, and is overlain by 6 m of King Formation outwash gravel (Fig. 6.27).

The Fish Formation is geographically more extensive than the underlying Traveller and Baxter formations. (Map 1). It covers all preceding sediments and consists mainly of outwash gravel and ice-contact stratified sediments. Tills are rare and difficult to identify because glacial comminution of the highly resistant source rocks of the West Coast Range produces little silt and clay.

Deposition of the Fish Formation was probably approximately synchronous with the Governor Formation as suggested by the similar aggradation levels in profile J - J', (Fig. 4.2) and the geometry of the contact (Fig. 6.31). It is likely that the sediment flux from the closer Mt. Jukes ice source (4 km) would have arrived before that of the King Glacier, the closest source of which is 25 km distant.

Outwash gravel extends southward to the Andrew Divide where up to 5 m of gravel rests on weathered Ordovician limestone. Similar outwash gravel occurs beyond the divide in the Andrew River catchment. The location of the gravel suggests that outwash streams from the King drainage system breached the divide and flowed into the Andrew River. The absence of till and large boulders in the area suggest that ice from the Mt. Jukes Glacier did not cross the divide (F. J. Baynes pers. comm. 1987).

4.8 The Governor Formation.

The Governor Formation, which was deposited by an advance of the King Glacier, is known only to occur in two exposures near Baxter Rivulet. It has no surface expression and it is not represented on the map showing the distribution of stratigraphic units.

The type section of the Governor Formation is an excavation 200 m south of the Kelly Basin Road bridge over the Governor River at G.R. 881303. The section consists of 4.3 m of crudely bedded outwash gravels that are iron stained, have multiple iron pans and are overlain by outwash gravels of the King Formation. The deposits of the King and Governor formations are easily distinguished on the basis of weathering of Jurassic dolerite clasts. The Governor Formation gravels have weathering rinds with mean values of 14.3 and 17.3 mm compared to 5.3 and 5.1 mm in the King Formation (Fig. 4.4).

Sediments of the Governor Formation have no surface expression because they have been eroded and buried by younger deposits. The absence of depositional morphology makes the limits of the ice advance difficult to determine. However, boulders up to 500 mm in diameter in Governor Formation gravels at Baxter Rivulet suggest that the ice terminated nearby, and that the advance was of a similar magnitude to the King Advance.

Although the absolute age of the formation is not known, estimates of its age in relation to the King Formation suggest that the sediments were deposited during a middle Pleistocene glaciation that occurred substantially before the Henty Glaciation (section 8.2).

4.9 The King Formation.

The King Formation was deposited by an extensive advance of the King Glacier. During this advance ice flowed 17 km down the King Valley and terminated near the bridge where a small moraine and outwash terrace are preserved. The formation consists mainly of terraces of outwash gravel with occasional lenses of buried till and a few low hills of ice-contact stratified deposits.

The type section of the King Formation crops out in a road batter at G.R. 884316, near the King River bridge. It consists of 18 m of poorly sorted outwash gravel resting on Eldon Group quartzite and is buried by up to 5 m of locally derived fan gravels and slope deposits. Till exposed

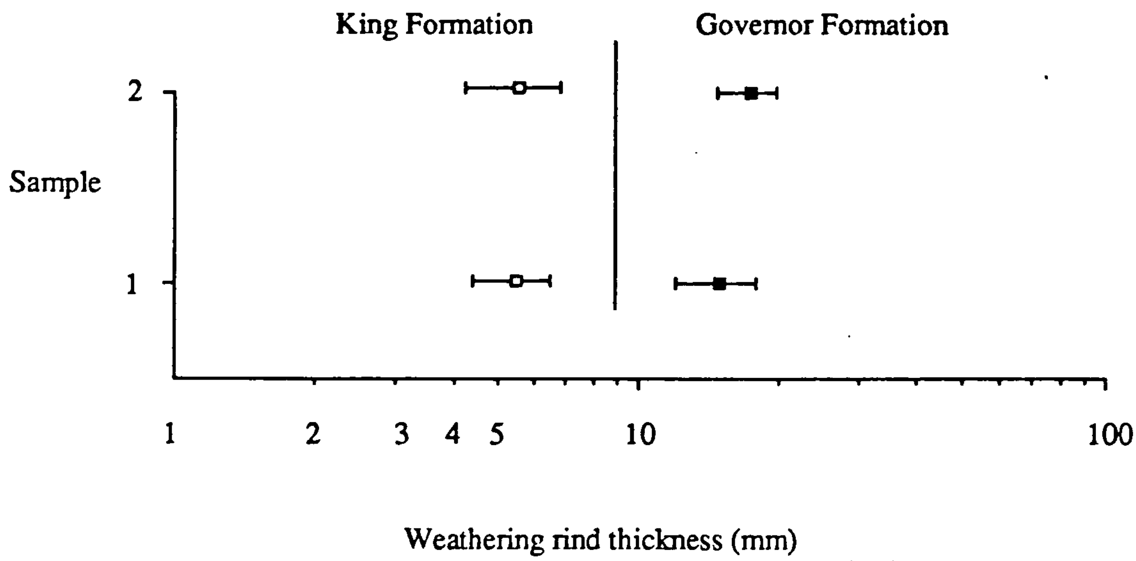


Figure 4.4 Differences in weathering rind thickness between the King and Governor and King formations. Each point represents the mean of 50 measurements and the bar represents ± 1 standard deviation.

in the type section, together with low ridges of ice-contact stratified sediments 200 m north of the type section, indicate that the extent of ice was similar to that of the Thureau advance (Fig. 2.6).

Profile F - F' (Fig. 4.2), shows the profile of the aggradation surface formed by deposits of the King Formation and profiles E - E', and G - G' show its upstream and downstream extent. The steeply-dipping aggradation surface reconstructed between sections G - G' and F - F', suggests that the surface is a remnant of an ice-proximal outwash fan.

Profile E - E', shows an aggradation surface 3.5 km. north of F - F'. Here the surface is underlain by 5 m of bouldery outwash gravel that occurs above the aggradation level of the Blackwood Formation. The terrace may represent a retreat stage or minor readvance of the King Glacier.

Although the limits of the King and Thureau advances in the King Valley are similar (Fig. 2.6), the evidence from the other distributary lobes clearly suggests that the Thureau advance was much more extensive. This apparent anomaly is probably due to the nature of the glacial system in which the watersheds of Linda and Comstock Creek controlled the thickness of ice in the King Valley. During the Thureau advance the divides were breached by ice that flowed into the Queen Valley. However, during advances when ice thicknesses were not sufficient to breach the divides the ice was concentrated in the King Valley. Thus, although the ice volumes of later glacier advances were less, their volumes and limits in the King Valley were similar. The Linda and Comstock watersheds acted as natural overflows that bled off ice from the King system during the Thureau advance.

4.10 The Nelson Formation.

The Nelson Formation is a 44 m sequence of laminated lake sediments that lie beneath the Blackwood Formation in the central part of the King Valley. The lake sediments rest on 10 m of gravel or till, a layer of cobbles and Devonian siltstone (Fig 6.57). The sediments are known from drill cores and large excavations at G.R. 896370, and occasional small exposures in the banks of the King River. The formation has no surface expression and is not represented on the stratigraphic map.

The sequence of laminated silts and muds accumulated at the bottom of a deep lake. The lake probably formed behind the King Formation terminal moraine which dammed the lower King Valley near the southwestern edge of the Thureau Hills (Fig. 2.1). The absence of dropstones and other glacial sediments throughout the sequence suggests that it accumulated away from direct glacial influences.

Some of the sediments were processed to extract pollen. Few grains were observed though there was substantial organic debris in the processed samples (G. van de Geer pers. comm. 1987). While the climatic conditions at the time of deposition are uncertain, the sediments clearly represent a significant period of time during which ice was not present in this part of the valley. However, the lake may have been fed by meltwater from remote glaciers. Although the formation cannot be claimed to have been a climatic interstadial similar to the Baxter Interstadial, it represents a long period of ice absence from the valley and lies between the deposits of two ice advances. It is therefore defined as an interstadial that is recognised on a lithostratigraphic basis.

4.11 The Blackwood Formation.

The Blackwood Formation is the most extensive glacial formation in the King Valley (Map 1). Its recognition is based on two excavations and numerous drill holes through an end moraine and outwash surface. The type locality at G.R. 896370, is named after the local name for a logging area. Profile D - D', (Fig. 4.2), shows the steep aggradation surface and end moraine which extends as an arc across the valley floor (Map 1). The excavations show that the formation consists of 29 m of interbedded till and outwash gravel that overlie laminated silt of the Nelson Formation (Fig. 6.57).

In addition to the main outwash plain, several terraces are cut into older deposits downstream of the type section. In the lower King Valley substantial erosion and redeposition of the Thureau, Governor and King formations took place during the Blackwood advance.

Dating of the Blackwood Formation is uncertain. Approximately 3 km east of Mt. Owen, an organic deposit at G.R. 887382 has a basal date of 37,800 $^{+800}_{-700}$ yrs. B.P. (SUA 2469) (Colhoun and van de Geer 1987). The organic deposits overlie Blackwood Formation till but appear to be separated by a considerable unconformity. The ^{14}C assay is regarded as a minimum age because of humic acid contamination (E.A. Colhoun pers. comm. 1986) and the Blackwood Formation is regarded as being older than the limits of conventional ^{14}C dating.

Magnetic measurements from the lake sediments of the Nelson Formation show a normal polarity (M. Pollington pers. comm. 1987). This is consistent with the interpretation that it lies between two advances of the Henty Glaciation. The highly weathered nature of Blackwood Formation tills suggests that a long period of time separates it from the Chamouni and Dante formations. Estimates of the magnitude of this time gap made from weathering rinds on dolerite suggest it is of the order of 250 ka.

4.12 The Bull Formation.

The Bull Formation is named after its proximity to Bull Rivulet in the upper part of the valley (Fig. 1.1). The profile of the moraine and outwash surface is shown in profile C - C', (Fig.4.2). There is no type section because there is no exposure of the moraine or outwash surface. However, a shallow pit dug on the moraine crest exposed weathered, poorly sorted gravels of glacial origin. The geographic distribution of this formation is very limited because subsequent erosion has destroyed most of the outwash plain. Although the duration of the period separating the Blackwood and Bull formations cannot be assessed, their proximity and similar altitudes suggest that the Bull sediments may represent a minor ice advance during a general recession from the Blackwood advance.

4.13 The Chamouni Formation.

Outwash gravels of the Chamouni Formation form extensive terraces inset into older glacial deposits (Map 1). The upstream extent of these terraces is unclear but they seem to be buried by a fan at the confluence of the Eldon and South Eldon rivers. The formation is named after the Chamouni Valley, an alternative name for the Linda Valley (Fig. 2.1).

The type section is a road cutting at G.R. 896432, which exposes 3 m of fissile lodgement till on the edge of a terrace. The terrace can be traced for 5 km down the King Valley (section B - B', Fig. 4.2). Eight hundred metres south of the type section at G.R. 893428 an exposure in the terrace shows 4 m of grey laminated silts which grade upwards into sands and a well sorted pebble gravel. The sequence records the changing depositional conditions during the onset of the Chamouni advance. The distinctive grey laminated silts underlie all subsequently deposited sediments in the upper King Valley.

Drifted wood and leaves from the upper surface of the silts, where they are unconformably overlain by Holocene gravels, were dated at 48,700 $^{+2900}_{-2100}$ yrs. B.P. (SUA 2599). A pollen count from the sediments from which the wood was taken indicates that the environment was not forested and probably had an alpine to subalpine vegetation (Table 4.3).

The degree of weathering of both the till and outwash gravel are similar to that of the Dante Formation and very different from the preceding Blackwood and Bull formations. These considerations, together with the location of the deposits between the limits of the Dante and Blackwood advances, suggest it was deposited during the Margaret Glaciation (Table 4.1) and therefore represents a separate ice advance.

The date, which is regarded as a minimum age, suggests the sediments are likely to have been deposited during the earlier part of the Margaret Glaciation. The age of these deposits is discussed

Table 4.3 Pollen from the Chamouni Formation.

TAXA		PERCENT
Temperate rainforest trees		**6.6**
Largarostrobos franklinii		2.9
Nothofagus cunninghamii		2.9
Phyllocladus aspleniifolius		0.8
Other trees and ferns	**5.6**	
Dicksonia antarctica		0.3
Eucalyptus		5.3
Small trees and shrubs	**12.4**	
Casuarina		0.5
Coprosma		1.3
Epacridaceae T-type		7.5
Labiatae		0.3
Letospermum		0.5
Orites		0.5
Papilionatae		0.5
Telopea		0.3
Tasmannia lanceolata		0.5
Pimelea		0.5
Alpine-Subalpine trees shrubs	**3.6**	
Athrotaxis-Diselma archeri type		0.3
Microcachrys tetragona		0.3
Microstrobos niphophilus		2.7
Nothofagus gunnii		0.3
Herbs	**71.7**	
Chenopodiaceae		0.3
Compositae Tubuliflorae		14.5
Cruciferae		0.3
Donatia novae-zelandiae		1.0
Gentianella		0.5
Gramineae		18.2
Gunnera		1.3
Haloragis		2.7
Labiatae		0.3
Liliaceae		0.3
Plantago		2.1
Polygonaceae		0.5
Ranunculus		4.5
Scrophulariaaceae		0.5
Umbelliferae		20.1
Hydrocotyle		0.3
Oreomyrrhis		4.3
Total pollen and spores in sum = 373		
OUTSIDE PERCENTAGE SUM		
Sedges	**8.5**	
Cyperaceae		7.5
Restionaceae		1.0
Aquatics	**0.8**	
Isoetes		0.3
Myriophyllum		0.5
Fern spores	**2.9**	
Gleichenia		0.5
Lycopodium fastigiatum		0.3
L. scariosum		0.3
Other trilete spores		0.8
Histiopteris		0.3
Other monolete spores		1.0

further in section 8.2.

4.14 The Dante Formation.

The Dante Formation is named after Dante Rivulet which flows from the Tyndall Plateau into the King River (Fig. 2.1). The type section at G.R. 902456, was described by Kiernan (1980). It consists of outwash gravel overlain by a palaeosol that in turn underlies outwash sediments of the Dante advance (Fig. 6.75). Wood 10 cm from the base of the Dante outwash was dated at 18,800 ± 550 yrs. B.P. (ANU 2533). Pollen from the palaeosol records an alpine herbfield-bog mosaic and contains a macrofossil of the alpine cushion plant *Donatia novae-zelandiae* (Gibson, Kiernan and Macphail 1987). The *Donatia* was dated at 21,180 ± 370 yrs. B.P. (SUA 2154) and drifted twigs stratigraphically below the *Donatia* were dated at 20,100 ± 470 yrs. B.P. (SUA 2155) (Gibson *et al.* 1987). Although Kiernan (1980) and Gibson *et al.* (1987) describe the lower outwash gravel as part of the Comstock Glaciation (= Henty Glaciation, Table 8.2, section 8.3) it is probably part of the Chamouni Formation described above and belongs to the Margaret Glacial Stage.

Kiernan (1980, 1983a) defined this section as representing the last glacial maximum in the central West Coast Range. The Dante Rivulet section has subsequently become the unofficial type site for the maximum of the Last Glaciation in Tasmania (Colhoun 1985a, 1985b, Colhoun and van de Geer 1986).

Profile A - A', Fig. 4.2, shows the small, steep Dante outwash fan and its relationship to the Chamouni Formation outwash surface. Unfortunately no section showed the contact between the Dante and Chamouni formations. A similar though larger fan with an end moraine at the apex occurs in the South Eldon River near its confluence with the King River.

4.15 The Long Marsh Formation.

Holocene sediments in the King Valley consist mainly of low-lying river terraces which commonly consist of about 1.5 m of buff coloured overbank silty sands resting on up to 3m of fluvial gravels. The gravels have been reworked from older glacial sediments which they overlie. Holocene terraces are mainly limited to the upper part of the King Valley near the Lyell Highway. Smaller areas of fluvially worked gravels occur also in the Nelson River and in the lower King Valley (Map 1, back pocket).

The type section of the Long Marsh Formation is a 3.5 m exposure in a terrace on the left bank of the King River at G.R. 892378. Core wood from a 30 cm log at 3.5 m depth was dated at 12,250 ± 90 (SUA 2415) suggesting that by 12 ka. B.P. the climate of the upper King Valley had sufficiently ameliorated to support the growth of large forest trees.

Wood from a low-lying terrace in the Nelson River was dated at 2150 ± 90 yrs. B.P. (SUA 2487) showing that recent reworking of the gravels has occurred.

In the central part of the King Valley wood from surface peat that overlies a gravelly slope deposit was dated at 2520 ± 80 yrs. B.P. suggesting that minor slope stability has occurred during the Holocene. This is also supported by evidence from Newall Creek where a thin bed of slope deposits occurs within the Holocene part of a deposit that has been analysed for pollen (van de Geer, Fitzsimons and Colhoun 1988 in press).

SUMMARY.

- Combined morphostratigraphic, lithostratigraphic and biostratigraphic classification and mapping of the King Valley has provided the most comprehensive knowledge of the Quaternary stratigraphy of Tasmania.

- The stratigraphy includes thirteen formations. Nine of the formations were deposited during ice advances from four glaciations. The extent of deposits associated with each ice advance is summarised in a map (Map 1, back pocket).

- The most extensive ice advance occurred during the early Pleistocene Linda Glaciation.

- The Linda Glaciation was followed by the deposition of organic sediments. These sediments record the successional development of temperate rainforest during the Regency Interglacial.

- The Governor Glaciation consists of two known ice advances the deposits of which are separated by organic sediments of the Baxter Interstadial.

- The Henty Glaciation consisted of three known ice advances. The deposits of the two older advances are separated by sediments of the Nelson Interstadial which records a long period of lacustrine deposition.

- The last or Margaret Glaciation consisted of two ice advances. The Dante advance occurred after 19 ka. and the Chamouni advance appears to predate 48 ka.

CHAPTER 5.

WEATHERING RINDS AS A RELATIVE DATING METHOD.

5.1 Introduction.

The purpose of this chapter is to examine the utility of weathering rinds as a relative dating method for the glacial deposits of the King Valley. The accuracy of the method can be assessed by comparing the weathering data from the King Valley with the stratigraphy described in Chapter 4. This is made possible because the stratigraphy was developed largely without the use of weathering rind data. The main exception to this was where comparisons were made between superimposed sedimentary units such as the Governor and King formations.

This chapter outlines the basis of the use of weathering rinds as a dating method. It presents the results from the application of the technique to the King Valley and contrasts them with the results of Kiernan (1983a) for the northern part of the study area. It also disuses the strengths and limitations of the technique as a dating method and as a means of correlating glacial deposits.

5.2 Factors affecting weathering rind thickness.

Several studies have used weathering rind thickness as a relative dating method for glacial deposits (Porter 1975, Chinn 1981, Colman and Pierce 1981). The use of the method is based on the assumption that rind thickness is age dependant and can be used to deliniate episodes of deposition (Burke and Birkland 1979).

Colman and Pierce (1981) use Jenny's (1941) model of soil formation to describe the processes of weathering in which the major factors of soil formation are climate, vegetation, relief, parent material and time (Fig. 5.1).

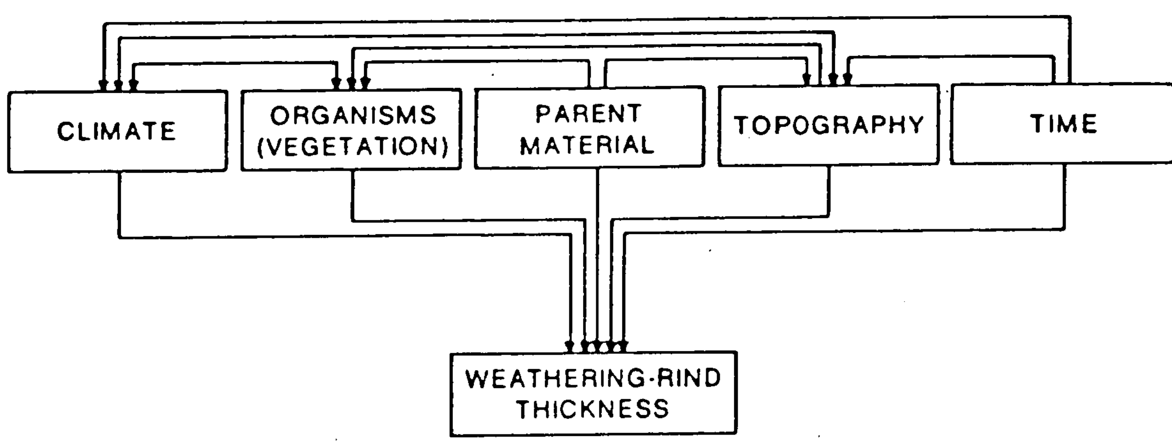


Fig. 5.1. Interrelations between the factors affecting weathering rind thickness (from Colman and Pierce 1981).

Climate is a major influence on weathering processes. The most important variables are probably temperature and precipitation. Colman and Pierce (1981) suggest that changes in climate over short distances can have an important effect on rind thickness. The main differences they observed were in relation to changes with altitude, soil moisture and precipitation.

The climate of Tasmania has strong precipitation and temperature gradients. The major rainfall gradient is from the perhumid west with a mean annual precipitation up to 3,600 mm at Lake Margaret to the subhumid east with mean annual precipitation down to 720 mm at Oatlands (Gentilli 1972). Within the study area there is little variation in precipitation but temperature decreases with increasing altitude from 11°C at 200 m at Queenstown to about 6°C at 1000m.

The role of vegetation as a soil forming factor is through the control of soil moisture, soil pH and soil erosion. Studies of weathering rinds in the United States have shown that there is little or no difference in weathering rinds developing under grass and forest vegetation (ibid p. 13). However, within the study area and in Tasmania as a whole, vegetation may have a significant effect on weathering. Kiernan (1985) noted that weathering rinds developed under buttongrass tended to be slightly thicker and to develop more rapidly than under open sclerophyll forests.

Although the measurement of rinds is commonly restricted to specific types of igneous rocks, there is frequently substantial lithologic and mineralogic variability within such rocks. In Tasmania the measurement of weathering rinds has been restricted to medium grained Jurassic dolerite. However, the considerable mineralogical and grain size differentiation in the sills and dykes may produce a variable response of individual clasts to weathering (Kiernan 1985). To what extent this influences the measurement of rinds for dating is not known.

The topography of the site affects the processes of weathering through site drainage, and erosion and burial of the weathering profile. Poor drainage may dramatically slow weathering by not allowing oxidation of minerals or the removal of solutes. Erosion of the weathering profile means that weathering of the sediments is not a true indicator of age. Burial of sediments results in elevation of the weathering front which is the depth to which weathering occurs. Depending on the location of the deposit, the older it is the more likely it is to be affected by erosion and burial.

There is a temporal aspect to all of the factors described above because they are all likely to change through time. In using weathering rinds as a dating method the most important temporal changes in the weathering factors are those that that do not affect all sites in the same way. The most important factors that may have such an effect may be due to localised erosion and burial of sediments. The effect of localised disturbance of weathering processes in the use of weathering as a relative dating method is discussed below.

5.3 Evidence from the King Valley.

The use of weathering rind thickness and other weathering index data has become an accepted method of relatively dating glacial sequences in Tasmania and considerable success has been claimed (Kiernan 1983a, 1983, 1985; Augustinus 1982; Augustinus and Colhoun 1987).

The thickness of weathering rinds on Jurassic dolerite clasts in the King Valley can be compared to the glacial stratigraphy defined by other methods (Chapter 4). From this comparison some assessment of the efficacy of rind measurement techniques can be made.

Fifty measurements were made at each site in accordance with the methods used by Colman and Pierce (1981) and Kiernan (1980, 1983a, 1985). Rinds were measured to the nearest 0.5 mm on stones sampled from about 1 m depth in the B horizons of soils. Each stone was either split or a caliper inserted through the rind to fresh rock. Multiple measurements were made on each clast and averaged to give the rind thickness.

Comparison of these data with those of Kiernan (1983a) for the same area (Figs. 5.2 and 5.3) shows there are major differences between the two data sets. Kiernan's data describes an apparently simple situation of three ages groups of weathering rinds thickenss in the deposits that

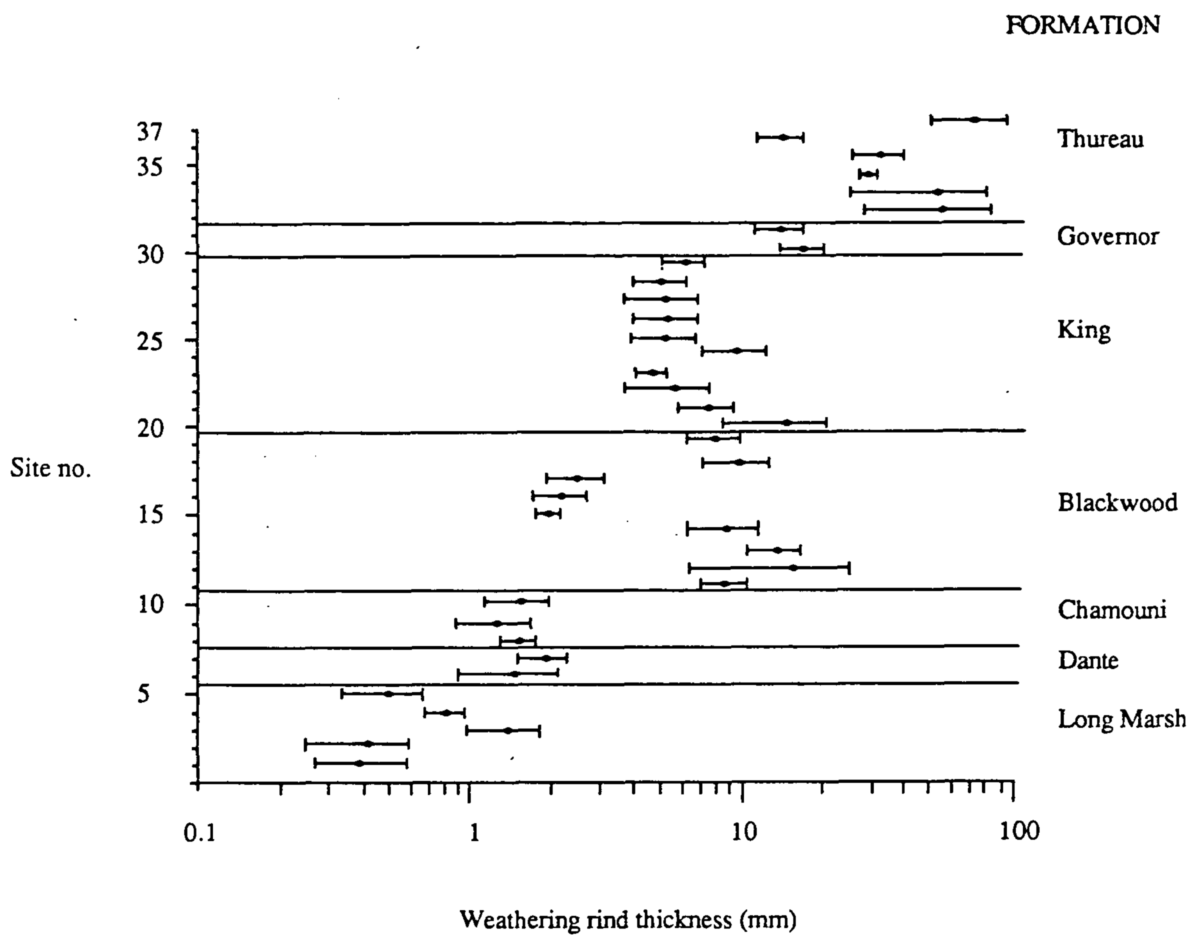


Figure 5.2 Summary of weathering rind thickness measurements from the different formations. Each point represents the mean of 50 measurements and the bar represents ± 1 standard deviation. Site locations are given in Appendix 2.

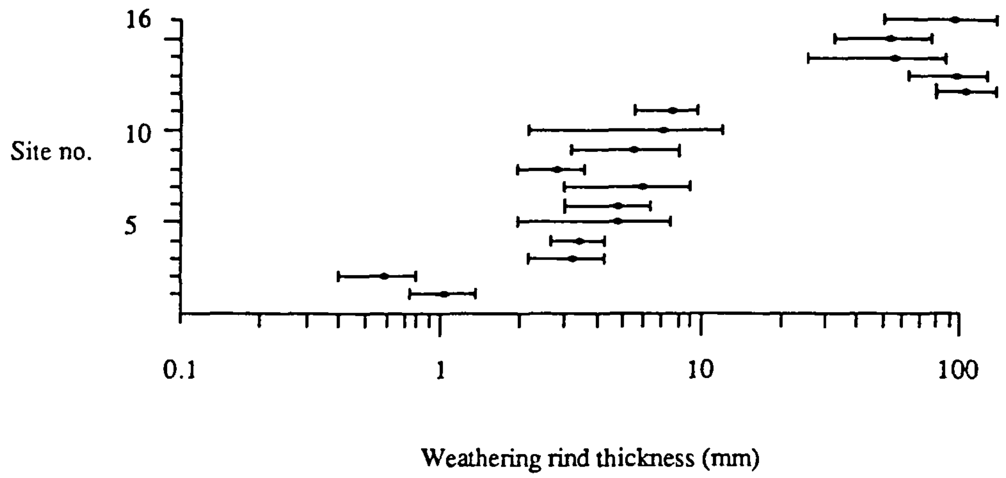


Fig. 5.3. Kiernan's (1983b) weathering data for the northern part of the King Valley. Each point represents the mean of about 20 measurements and the bar represents ± 1 standard deviation.

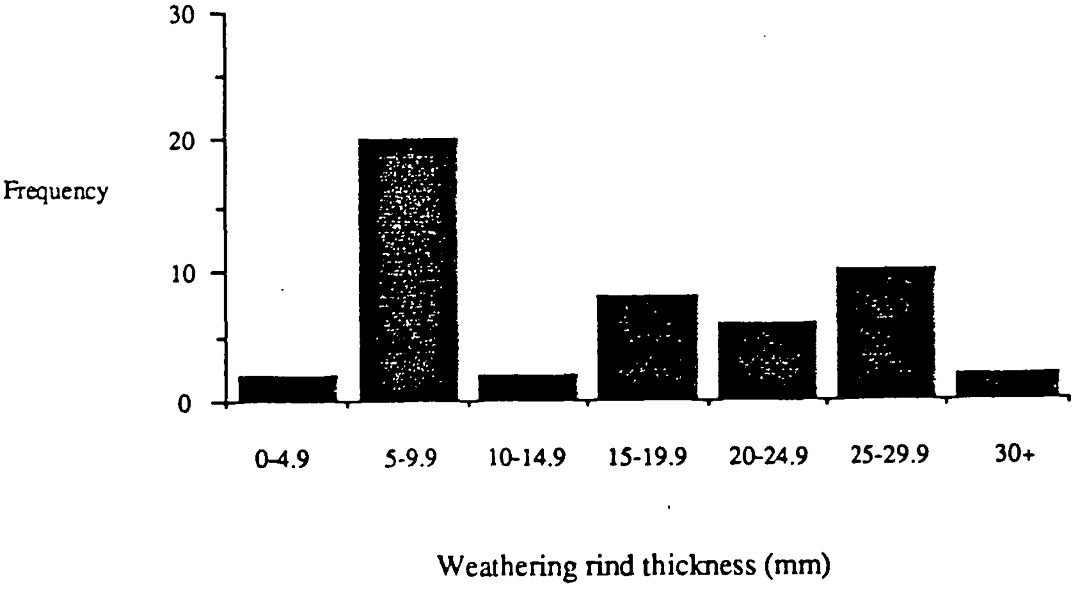


Fig. 5.4. Distribution of weathering rind measurements from site 12, Fig. 5.2.

are separated by order of magnitude differences. He interpreted the evidence as indicating three glacial stages separated by interglacial periods. Although Kiernan used a smaller sample size than this study (c20 v's. 50) there was generally good agreement between mean measurements of common sites.

Patterns within the data set from this study highlight many of the problems of using weathering rinds as a relative dating method in Tasmania. Although mean thickness of weathering rinds show a general increase with increasing relative age of formations (Fig. 5.2), there are several exceptions to the trend. These are:

1. an anomalously high standard deviation in the measurements from site 12;
2. weathering rinds from the Blackwood Formation are more variable and are thicker at many sites than those of the older King Formation;
3. a wide range in the thickness of weathering rinds of Thureau Formation sediments.

The high standard deviation of site 12 is caused by the weathering rind sample being multimodal (Fig. 5.4). Although there are several alternative explanations for this including variation in grain size and/or mineralogy, density of the dolerite and within site differences in the weathering rate, there is no evidence to suggest a specific cause. The geometry and structure of the sediments suggests that site 12 is a deformation till deposited during the Thureau advance that was sheared and eroded by a subsequent ice advance (section 6.8.3). This, together with the multimodal weathering rind distribution (Fig. 5.4), suggests that some of the clasts were preweathered, i.e. they did not completely lose their weathering rinds when the deposit was eroded and redeposited.

The differences in thickness of weathering rinds in tills and outwash gravels of the Thureau Formation suggest there was a difference in weathering rate between the different sediments. This is not unexpected, weathering rinds on outwash gravels are frequently much thinner than those on tills. Colman and Pierce (1981) demonstrate that rinds on outwash gravels are thinner than on tills by a factor of 0.8. On Jurassic dolerite in Tasmania Kiernan (1985) found the factor to be 0.89.

In contrast, the factor in the Thureau Formation measurements is 0.5. Although the reason for this difference between outwash gravel and till is not clear, it may be related the matrix grain size of the deposits. That is, deposits with finer grain sizes may experience higher rates of chemical decomposition. However, the differences within the Thureau Formation are not limited to outwash gravels, they also occur within tills.

There are two groups of rind thicknesses in the Thureau Formation tills, one with a mean near 30 mm and the other with a mean of about 60mm (Figs. 5.2 and 6.18). Because sites 25 and 26 are 4 m apart in the same section these differences suggest that small, within section differences in weathering rates can occur. This, together with the considerable overlap of almost all measurements and the reversal of the expected differences between the King and Blackwood Formations, suggests that weathering rates have changed since the deposits formed. Because all the deposits occur within 20 km of each other, climatic change can be considered to have had the same effect on all deposits. The old age and complex post-depositional history of deposits in the lower King Valley suggest that changing topography is a likely cause of the changes in weathering rates.

Several topographic factors may change over time and introduce errors into the assumption that time is the primary correlate of weathering rind thickness. Observation of sediments in Tasmania suggest three site factors can significantly reduce theoretical maximum weathering values. These are:

1. erosion of sediments;
2. burial of sediments;
3. high water tables.

Erosion of sediments can completely remove the weathering profile at any time (Fig. 5.5a) and is more likely to occur over longer periods of time and in areas that experience episodic climatic change than in relatively young deposits that have not experienced as many cycles of climatic

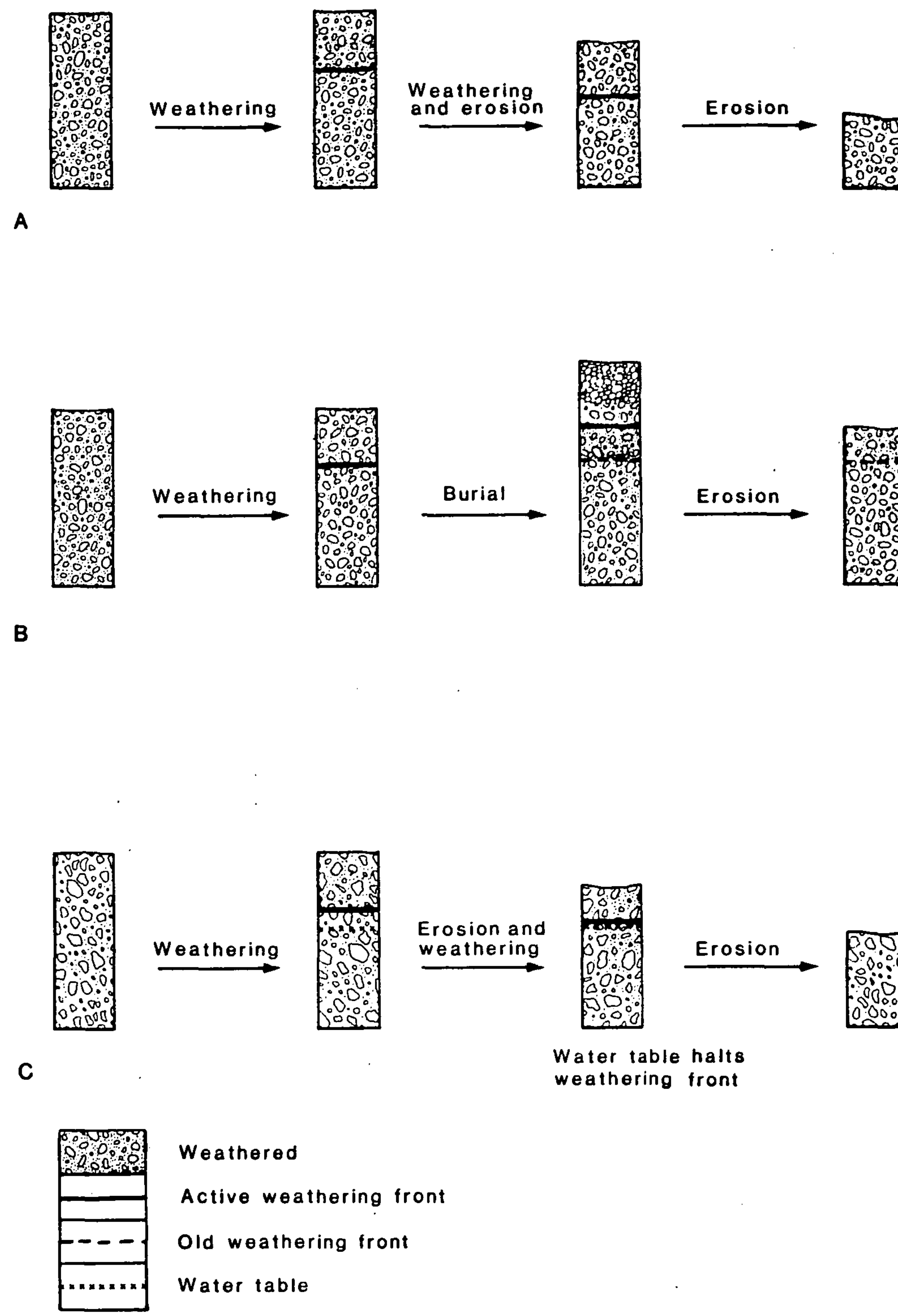


Fig. 5.5. Models of the effect of changing topography on the development of weathering fronts and weathering rates. A, Erosion. B, Burial. C, High water tables.

change. Colman and Pierce (1981) attribute considerable scatter in weathering rinds of tills near Mt. Rainer to erosion of many of the sites. They concede that despite attempts to avoid eroded sites, minor differences in topography may have appreciable effects on the thickness of weathering rinds, and that uneroded sites older than about 100 ka. probably do not exist.

Burial of sediments causes the weathering front to be lifted (Fig. 5.5b). The weathered sediments below the weathering front then experience a lower rate of weathering. Colman and Pierce (1981) noted that burial of glacial deposits by loess greater than 100 cm thick significantly slowed weathering rates.

Drainage can influence weathering of sediments by halting the downward extension of weathering fronts (Fig. 5.5c). Such a situation is likely to occur where high water tables inhibit the removal of solutes and therefore slow the rate of chemical weathering. An example of this situation is known from a site near Derwent Bridge in central Tasmania where a highly weathered till overlies an unweathered till in a recently drained quarry. It is quite clear that if erosion strips the weathered deposit and brings the unweathered till to the surface, measurement of weathering rinds will give an inaccurate indication of relative age.

The three objections outlined above mean that the accuracy of weathering rind measurements as relative age indicators decreases with age and with increases in the complexity of erosion history. It has been suggested that a possible means of removing error introduced by the erosion and burial of sites is to restrict sample sites to uneroded, primary depositional landforms as was done by Kiernan (1985). However, while Kiernan claimed to be sampling uneroded landforms, other criteria he used for relative dating included width of moraine crests, moraine slope angles and moraine relief which are all based on degradation of moraines by erosion over time (*ibid* p.77). This suggests that while the problem of erosion has been known, its implications for the use of weathering data have either not been recognised, underestimated or discounted. It is the experience of the author that uneroded primary landforms older than middle Pleistocene in age are extremely rare if not absent in Tasmania. This observation is supported by a study of soil profile development on glacial sediments of different ages in western Tasmania which concluded that all soil profiles on glacial deposits of pre Last Glaciation age are eroded (Hammond 1985).

5.4 Implications for the use of weathering rinds as a dating method in Tasmania.

The studies on weathering rinds on Jurassic dolerite in Tasmania are unique because of the antiquity of the deposits studied (>730 ka.) and because of the great thickness of the rinds (>100 mm).

While studies of weathering rinds as relative age indicators on deposits younger than 200 ka. old have been successful in the U.S.A. and New Zealand (Porter 1975, 1976; Colman and Pierce 1981; Chinn 1981), the same cannot be claimed for studies of older deposits.

The utility of the technique in deposits older than about 200 ka. and in areas with a complex and changing post-depositional history is limited because rind thickness is no longer primarily a function of time. Because most glacial sediments in the King Valley are older than 100 ka. and have a complex post-depositional history the use of weathering rinds as a dating technique is not altogether appropriate. The major exception to this is the use of weathering rinds to directly compare deposits overlying each other. This has been demonstrated at Baxter Rivulet where assessment of the differences in mean rind thickness permits estimation of the time interval between deposition of the King and Governor formations (see section 6.5).

Whilst not disagreeing with Kiernan's thesis that weathering rinds provide evidence for three glaciations (Kiernan 1983a), the errors in the technique on deposits older than about 200 ka. are such that one cannot be sure that the age of any one deposit is accurately reflected by weathering rind thickness.

Collection of the weathering rind data from the King Valley has shown that weathering data will not separate every glacial formation. The main cause of the inconsistencies appears to be due to the erosion, burial and drainage history of particular sites. When these sources of error are combined with the large regional gradients in precipitation and temperature across Tasmania correlation of glacial sediments based upon weathering rind data must be treated with caution. Use of the measurements has to take into account not only variations in in the factors governing weathering both spatially and temporally, but also site factors largely related to the history of the modification of the surface deposits.

In favour of the technique, it appears that the large age differences between deposits of the Margaret, Henty and Linda Glaciations are such that the weathering rind evidence provides a reasonable guide to approximate stratigraphic age. It is only as the division of glacial deposits has become finer with discovery of more complex stratigraphies that the technique has become less useful and more subject to the errors outlined above.

SUMMARY.

- The use of weathering rinds on Jurassic dolerite in Tasmania identifies three glaciations (Kiernan 1983a).

- Comparison between the complex stratigraphic sequence for the King Valley shows that weathering rinds do not necessarily indicate the relative ages of all the glacial deposits.

- The main sources of error appear to be due to the erosion, burial and fluctuation of groundwater levels in the sediments. Measurement of weathering rinds from deposits so altered do not give data that are truly age dependent within one set of regional climatic controls.

- The older the deposit the more likely it is to have been affected by the local site factors which may alter the results of the weathering analysis such as to render comparison of relative ages difficult and in many cases impossible.

CHAPTER 6.

SEDIMENTOLOGY.

6.1 Introduction.

Study of the glacial sediments in the King Valley provided the basic information for stratigraphic classification (Chapter 4) and for the reconstruction of elements of the King Glacier debris system (Chapter 7).

This chapter outlines the problems of environmental reconstruction and presents the data obtained from the major sections in the King Valley. The following chapter (Chapter 7) uses these results to consider the overall controls of the patterns and processes of glacial sediment genesis in the King Valley.

6.2 Reconstruction of sedimentary environments.

There are three different parts to the reconstruction of depositional environments, description of the sediments, interpretation of the processes of deposition, and detailed interpretation of the environment using both modern and ancient analogues (Anderton 1985).

Accurate descriptions of glacial sediments is essential because they form the basis of all subsequent interpretations and can limit the quality and detail of interpretation. Objectivity of description has been stressed in recent reviews of environmental interpretation because initial interpretations can prejudice an entire investigation Eyles *et al.* (1983). It is essentially for this reason that the term diamicton is used throughout this thesis and the term till is avoided unless an interpretation of a specific glacial origin is made.

Deciding on the amount and type of data recorded at each section was a constant problem in this investigation. The decision on the detail and type of information to record was a trade off between what information is considered essential in the interpretation and the time available. Fortunately, it was possible to revisit most of the sections in the King Valley if it was necessary to collect more information.

The most important sections in the King Valley were recorded in two dimensions to record the lateral variability of the sedimentary environments. This is particularly important in glacial environments which are characterised by abrupt lateral changes in sedimentary character and environment.

Data recorded in the King Valley are partly quantitative and partly qualitative. Qualitative aspects include mapping the geographic distribution of the deposits, studying the geometry of the deposits, examining the vertical sequence of sediments, the type of contact, the primary and secondary sedimentary structures, the colour, weathering, texture, sorting and the clast shape. Quantitative data, include, pebble fabric, particle size, weathering and stone count analyses as described in Chapter 3.

6.3 The Gormanston Formation.

Sediments of the Gormanston Formation are the oldest unconsolidated sediments known in the study area. Although their age is not accurately known, they antedate sediments of the Linda Glaciation and may be of late Tertiary age. Sediments of this type are known only from one section in the Linda Valley where they crop out on an eroded spur in a small gully.

The sediments consists of 12 m of bedded fluvial gravel and sands that dip at c 23° to the southwest towards a rock wall (Fig. 6.1). Although the contact is not seen, the sediments are

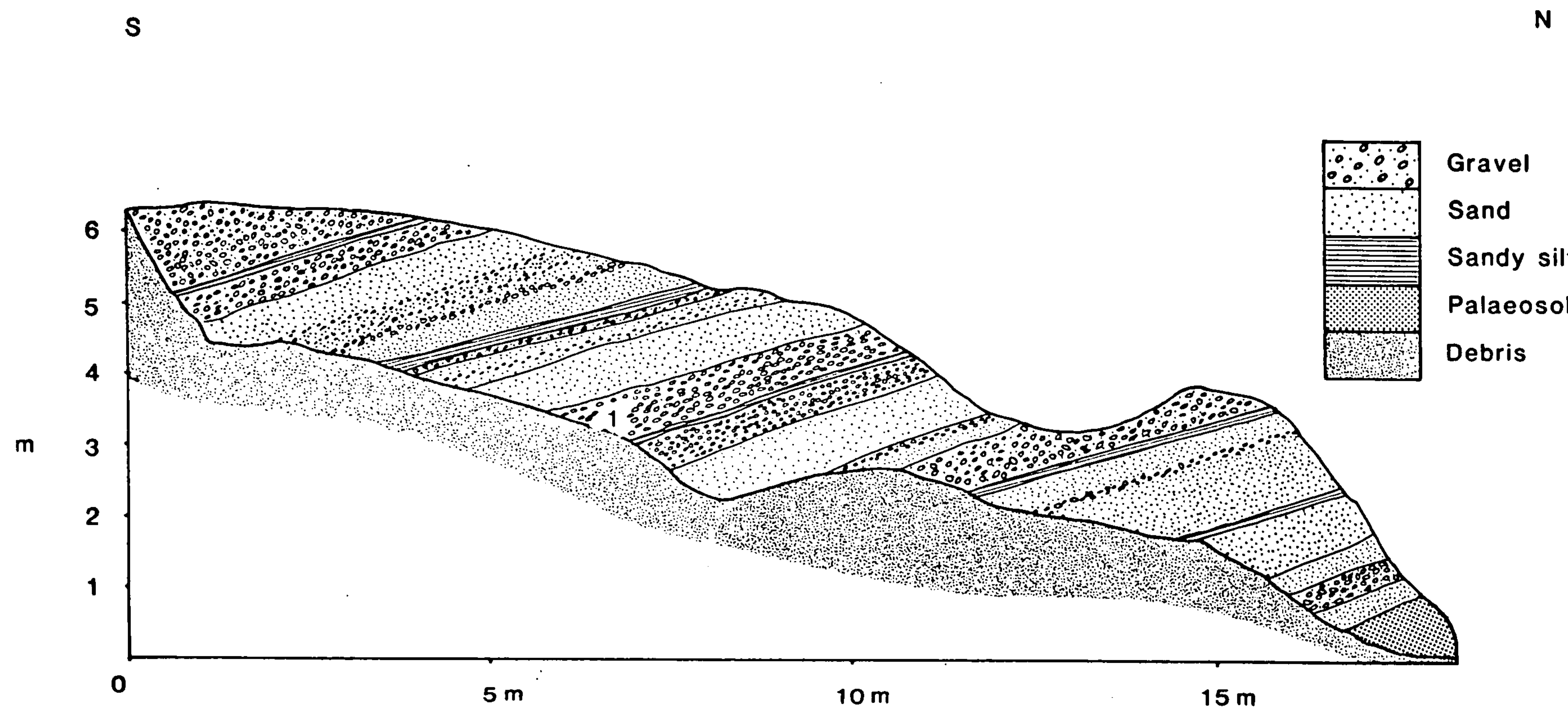


Fig. 6.1. Interbedded sands, gravels and silts of the Gormanston Formation.

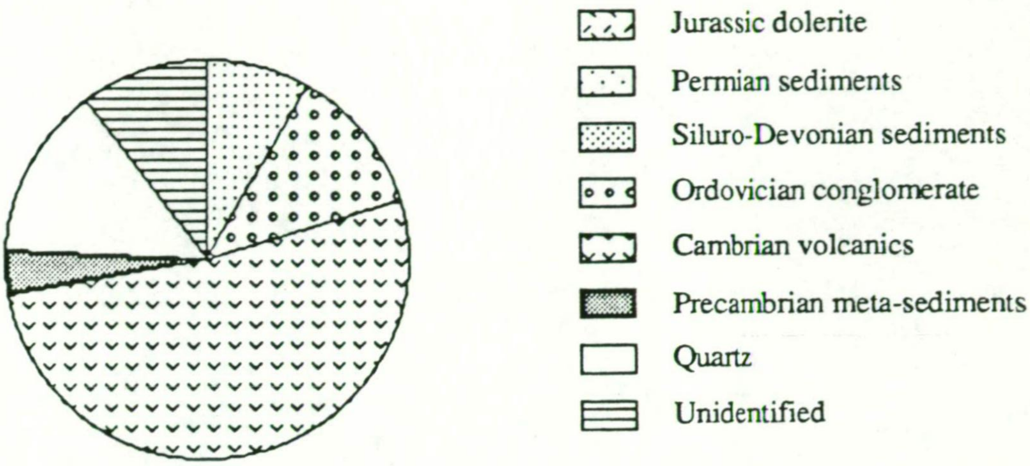


Fig. 6.2. Lithology of the Gormanston Formation sediments, location of sample site is shown on Fig. 6.1.

Fig. 6.3. Palaeosol underlying the Gormanston Formation.

inferred to lie beneath Thureau Formation glacigenic deposits which crop out above and on either side of the section.

Description.

The base of the section consists of a 1.5 m thick palaeosol that contains drifted wood and stumps of trees in growth position (Fig. 6.3). Pollen from the palaeosol consists of a wide range of forest taxa some of which have a Tertiary affinity (Table 8.1, section 8.2). The palaeosol contains abundant charcoal and is developed on coarse quartz sand.

The palaeosol is overlain by 40 cm of coarse sand that contains numerous wood fragments up to 2 mm in diameter derived from the palaeosol. The remainder of the sediments consist of layers of coarse massive sand interbedded with well-sorted pebble gravel and occasional thin layers of laminated sandy silt all of which dip between 15 and 23° (Fig. 6.1).

The lithology of the sediments suggests they are locally derived (Fig. 6.2) which contrasts with the high erratic component in the glacigenic sediments that crop out above and around the section. Pebbles in the gravels are well rounded while the granules and pebbles in the sands are angular and subangular suggesting that they have not traveled far.

Interpretation.

The provenance, texture and geometry of the sediments suggest they are quite different from the overlying and surrounding glacigenic sediments. Although earlier interpretations of this section suggested that it was at least partly of glacial origin (Kiernan 1980), the complete absence of erratic rocks suggests they were deposited prior to the deposition of overlying glacial sediments.

The geometry and apparent stratigraphic relationship of the sediments with Thureau Formation sediments suggests that the Gormanston Formation sediments are an eroded, non-glacial, possibly Tertiary, inlier within Thureau Formation glacial lake sediments.

The geometry, structure and particle size of the sediments suggest they accumulated as horizontal beds in a fluvial environment and were subsequently tilted or on the foreset beds of a delta. Although tilting of Tertiary sediments in western Tasmania has been recorded at Macquarie Harbour by Solomon (1962), the lack of other deformation structures suggests the sediments have not been disturbed by tectonic movements. The sequence probably accumulated on the foreset slope of a delta and the sediments were derived from the surrounding rock slopes.

6.4 The Thureau Formation.

Introduction.

Thureau Formation sediments are exposed in five principal areas, in the lower King Valley near the Governor River confluence, adjacent to the Thureau Hills, in the Linda Valley, in the Nelson River Valley and at Newall Creek. The most extensive sediments accumulated in ice contact lakes in the Linda and Nelson Valleys which were outside the limits of subsequent advances (Map 1). All other exposures have been subject to erosion by younger ice advances and are frequently buried by younger outwash gravel, fluvial deposits and slope deposits.

The type section.

The type section of the Thureau Formation is the northern most road cutting in a series of exposures of highly weathered till buried by slope deposits from the adjacent Thureau Hills. The exposure at G.R. 882353 is cut through a steep colluvial fan surface that buries a suite of glacial sediments. The underlying glacial sediments consist of massive till overlain by a series of flow tills, massive sands, a thin sediment flow and laminated lake sediments (Fig. 6.4). The whole sequence was penecontemporaneously deformed by the melting of ice on which it was deposited.

Description.

Unit 1.

The diamicton at the base of the section is coarse, massive, matrix-supported and contains small lenses of stratified sands and gravels. The lithology is dominated by erratic Jurassic dolerite and Permian sediment clasts which form a larger proportion than in the overlying sediments (Fig. 6.5).

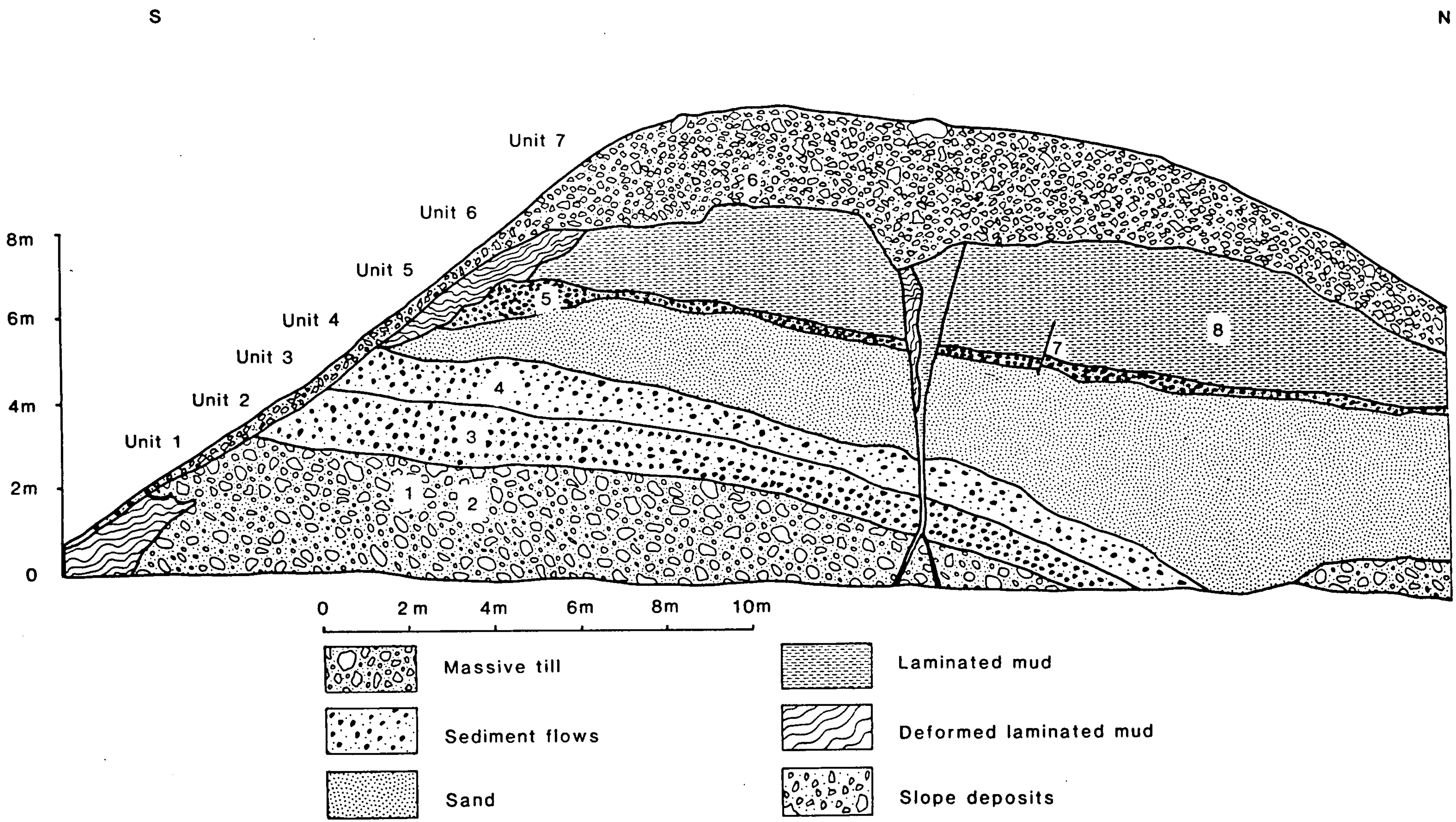


Fig. 6.4. Type section of the Thureau Formation. Massive till overlain by multiple sediment flows, massive and laminated sand, a sediment flow, laminated silts and slope deposits. Numbers refer to location of pebble counts and fabric analyses.

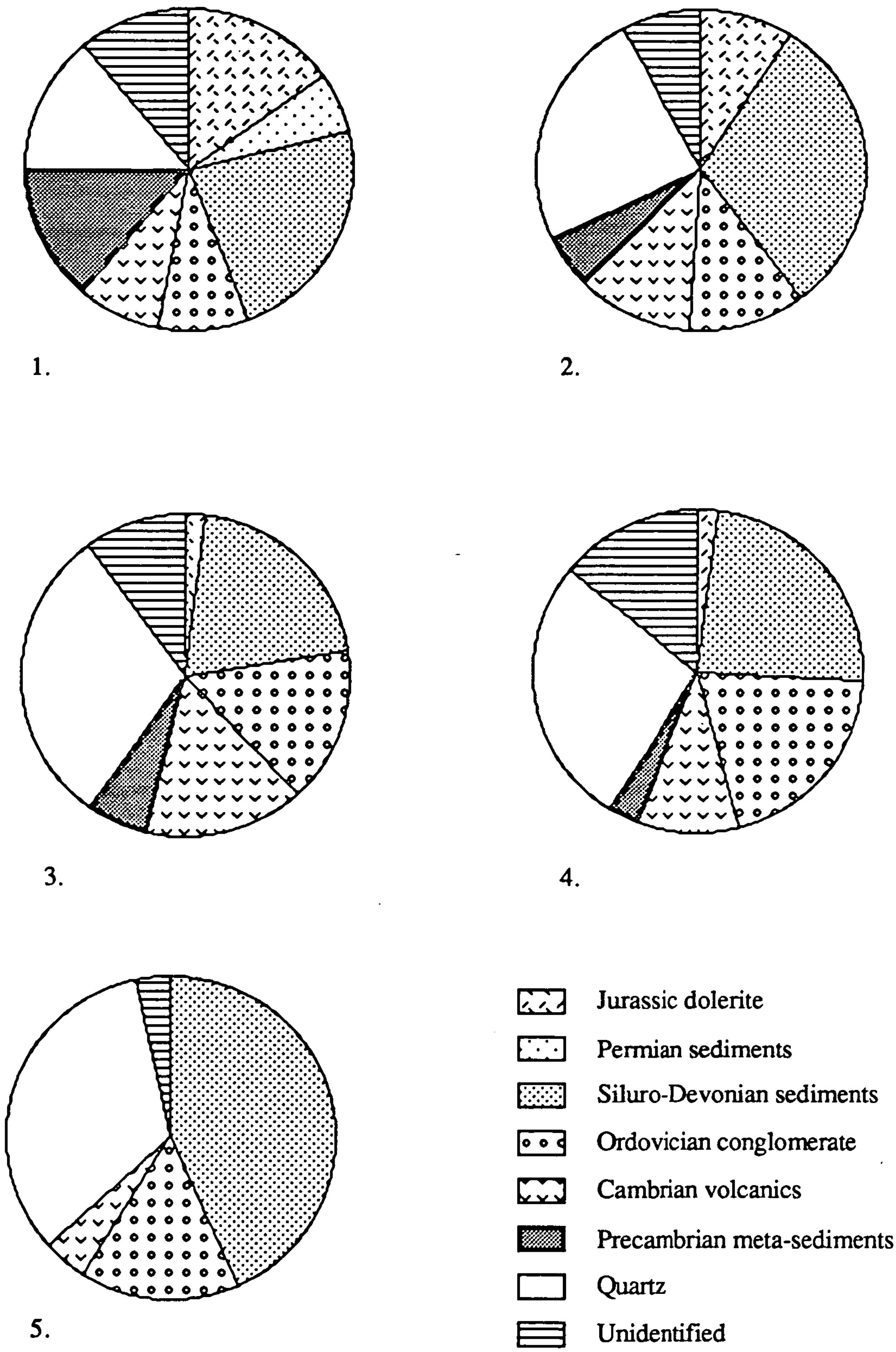


Fig. 6.5. Lithology of sediments exposed in the Thureau Formation type section. See Fig. 6.4 for sample locations.

Although the position and characteristics of the diamicton suggest a melt-out origin, because the pebble fabric is dispersed and more typical of sediment flows, (Fig. 6.6) it is difficult to interpret as a melt-out till. The dispersion of the pebble fabric is similar to the overlying sediment flows.

Unit 2.

This 1.2 m diamicton is unsorted, has occasional flow noses, and has rudimentary flow banding that dips north at 15°. Jurassic dolerite clasts form a significant part of the lithology of this deposit (Fig. 6.5), are highly weathered (Fig. 6.7b) and have weathering rinds with a mean thickness of 54.5 mm and standard deviation of 28.2 mm. The matrix is also highly weathered and bleached to a pale grey colour (Fig. 6.7). The pebble fabric has a weak concentration around 267° that is apparently unrelated to glacier flow (Fig. 6.6, fabric 3). The contact with the overlying sediment flow is sharp, and is marked by a scoured surface that is parallel to the flow banding and bedding of both diamictons.

Unit 3.

The diamicton is up to 1.3 m thick and consists of 4 layers of fine clast-supported pebbles with intervening layers of highly weathered and bleached, pebbly diamictons that appear to be sediment flows (Fig. 6.7). The pebble fabric is dispersed and appears to reflect the palaeoslope on which the sediment flows and meltwater deposits formed (Fig 6.6, fabric 4). The provenance of the flows is dominated by West Coast Range and local source rocks with rare pebbles of erratic Jurassic dolerite and Permian sediments (Fig. 6.5). The contact with the overlying sand is sharp and overlying sand appears to bury a depositional surface that dips north at between 10 and 16°.

Unit 4.

Overlying the diamicton is up to 3.5 m of moderately well sorted coarse white quartz sand. The top 1.5 m is horizontally laminated with thin beds of silty sand and the remainder is massive. A small exposure in a drain 10 m north of the section exposes 3 m of laminated fine sand and silty sand (Fig. 6.8). Although the relationship is unclear these sands are interpreted as a northward

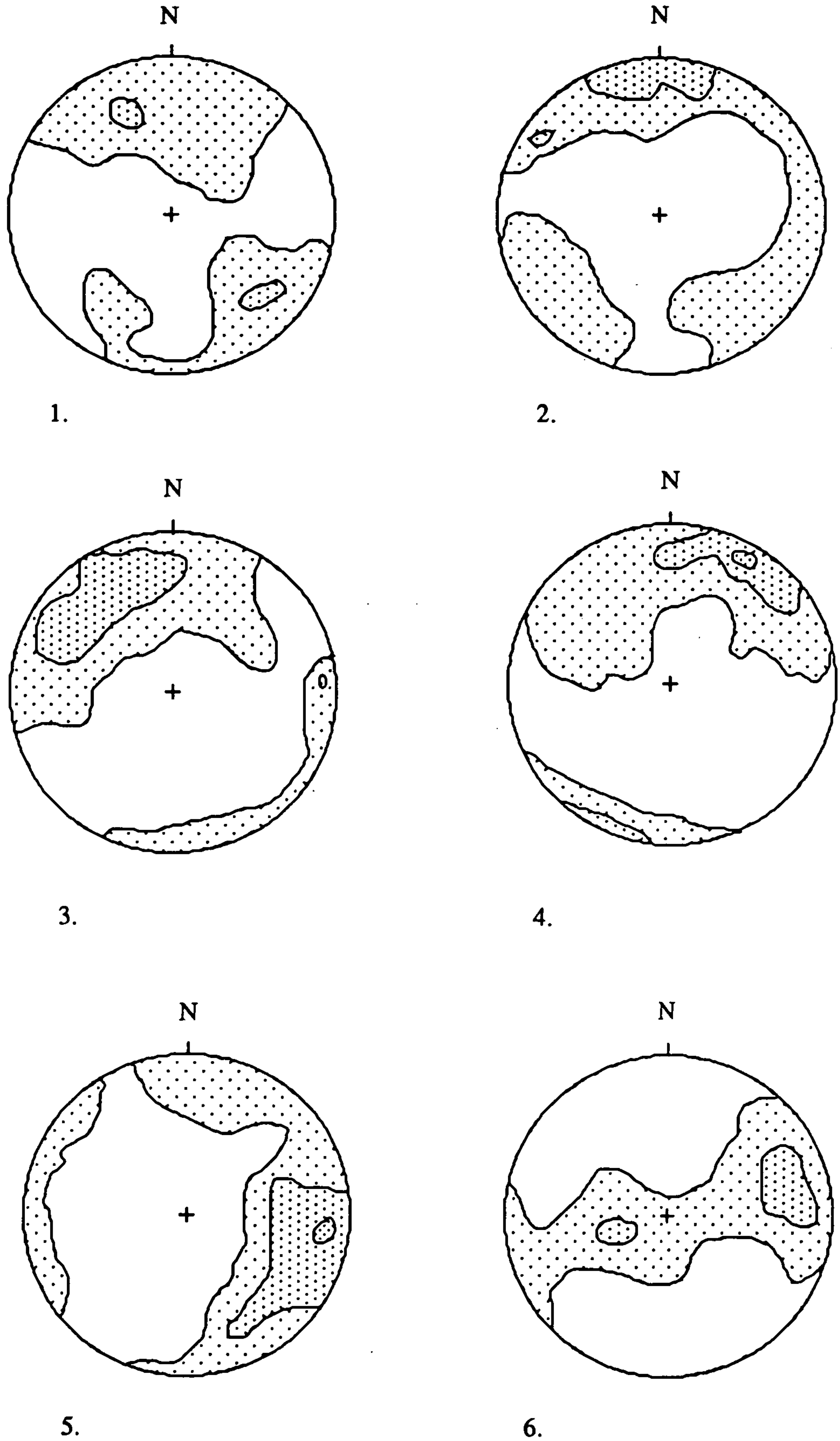
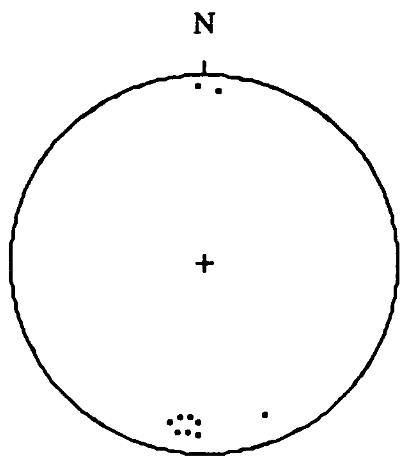
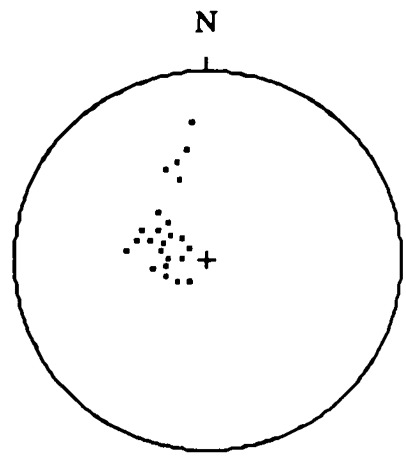


Fig. 6.6. Contoured equal area plots of pebble fabrics from the type section of the Thureau Formation. See Fig. 6.4 for sample locations. Contour interval is 2 standard deviations.



7. Fold dip direction



8. Faults

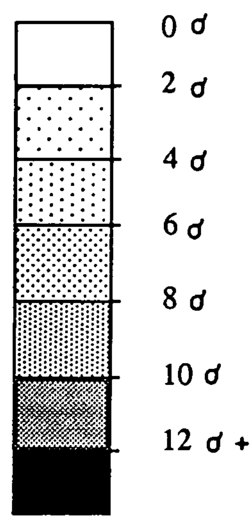
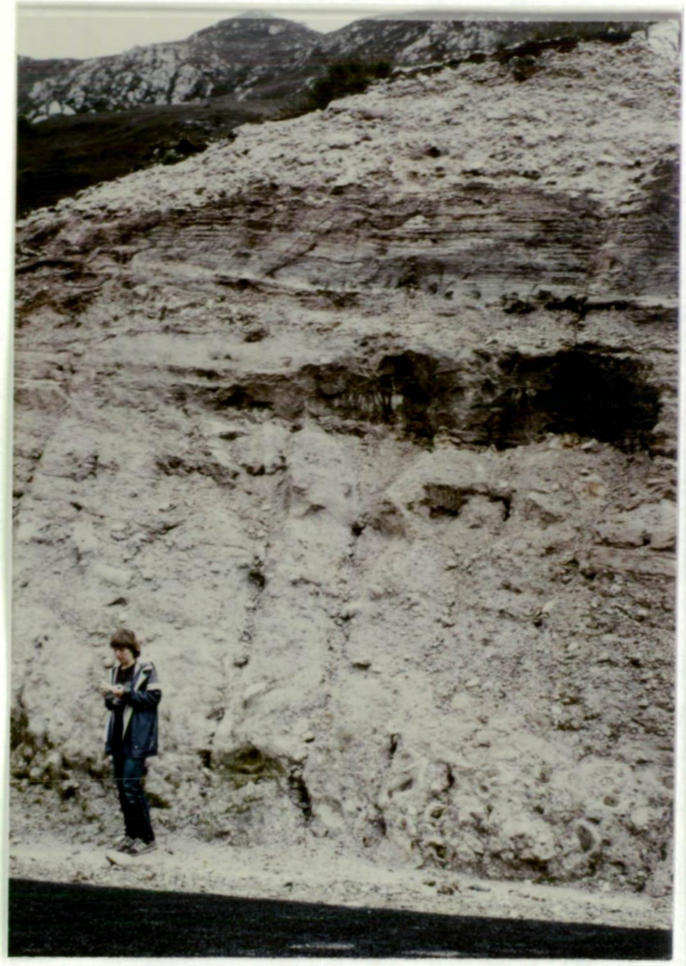


Fig. 6.6. (ctd.) Equal area plots of faults and folds in the Thureau Formation type section. See Fig. 6.4 for sample locations.



A

B

Fig. 6.7. The Thureau Formation type section. A, Interbedded sediment flows and sheet flow deposits. B, Highly weathered Jurassic dolerite cobbles below a sediment flow.

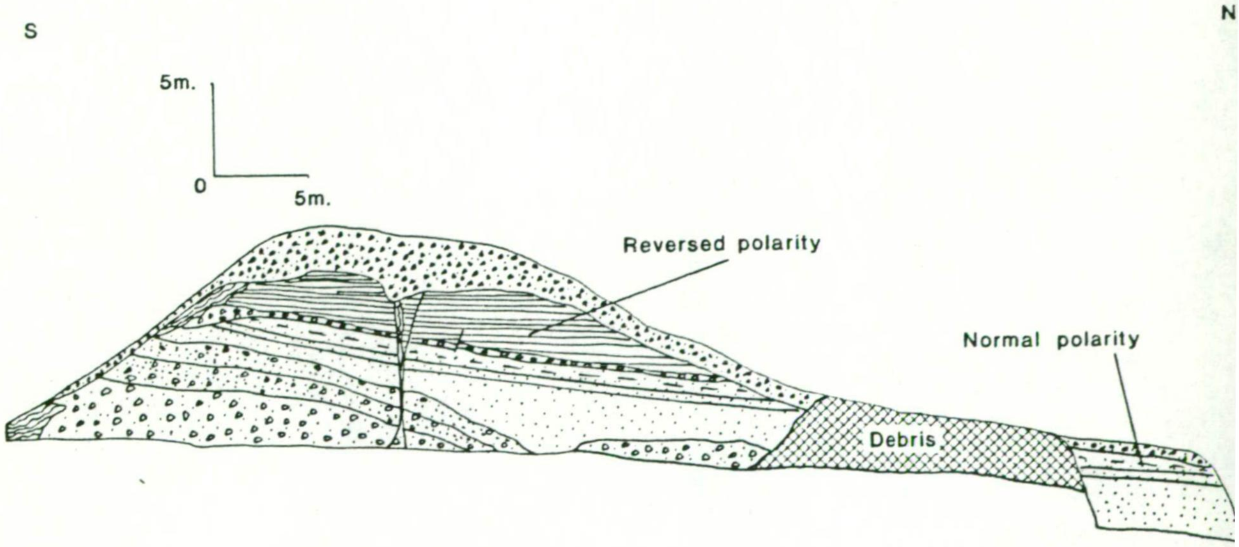


Fig. 6.8. Relationship between the type section and sediments that have a normal detrital remanent magnetisation.

thickening wedge of the sand in the main section. This interpretation is crucial to understanding the temporal relationships of the sections because the sands have a normal detrital remanent magnetisation while the overlying silts have a reversed polarity (Fig. 6.8). It is difficult to escape the interpretation that the section crosses a palaeomagnetic boundary, possibly from the Jaramillo normal event into the reversed part of the Matuyama Chron, a boundary known to occur at 910,000 yrs. BP (Bowen 1978). Unfortunately, because the field relationship of the two units is unclear the interpretation of the sediments crossing a palaeomagnetic boundary is uncertain. The contact between the sand and the overlying diamicton is sharp and eroded.

Unit 5.

The overlying diamicton is up to 1 m thick and consists of a thin bed of rounded pebbles in a matrix of grey silt similar to the overlying laminated silts. It dips to the north at 11°. It is structureless except for the occasional fault which extends from the overlying silt. The lithology of the diamicton consists largely of rocks with a local and West Coast Range provenance and rare erratic clasts (Fig. 6.5). The pebble fabric is weak and unrelated to the ice flow direction which was from north to south (Fig. 6.6). Numerous beds of the underlying sand are truncated by the diamicton which may be a thin sediment flow or ice-rafted diamicton deposited during the onset of lake deposition.

Unit 6.

Up to 3 m of dark green and dark grey and green laminated silts are draped over the thin diamicton. The dark green laminae are silts and the dark grey laminae are clayey silts according to the classification of Folk, Andrews and Lewis (1970), (Fig. 6.9). Individual laminae are well sorted and there appears to be minor upward-fining textural differentiation in the green laminae.

Numerous small-scale primary and secondary structures that include convolute lamination and reverse faults are common in the silts. The syndepositional nature of the convolute lamination is apparent from its occurrence in specific beds which are overlain by undisturbed beds (Fig. 6.10). The trend of fold axes of the convolute lamination are highly concentrated and dip northward

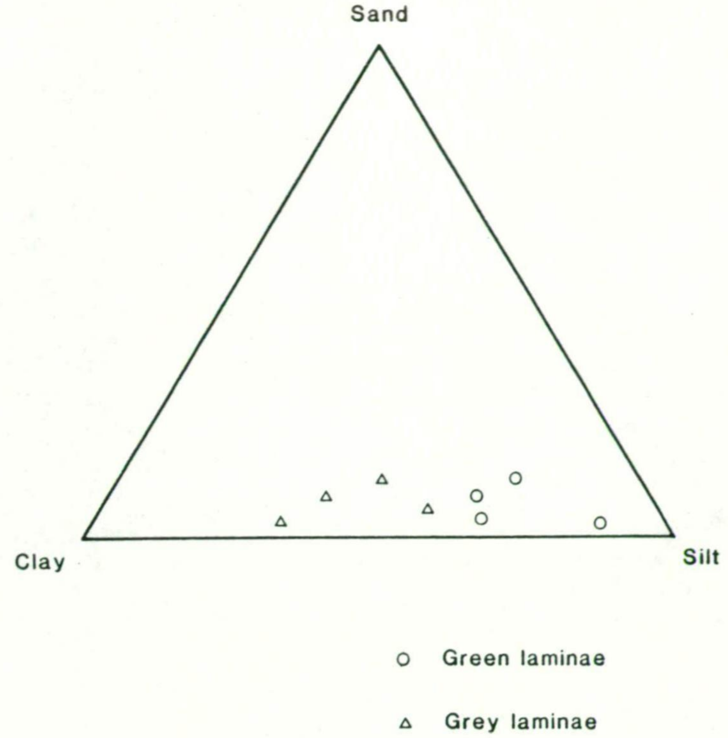


Fig. 6.9. Ternary diagram of the particle size of laminated silts of the Thureau Formation type section.

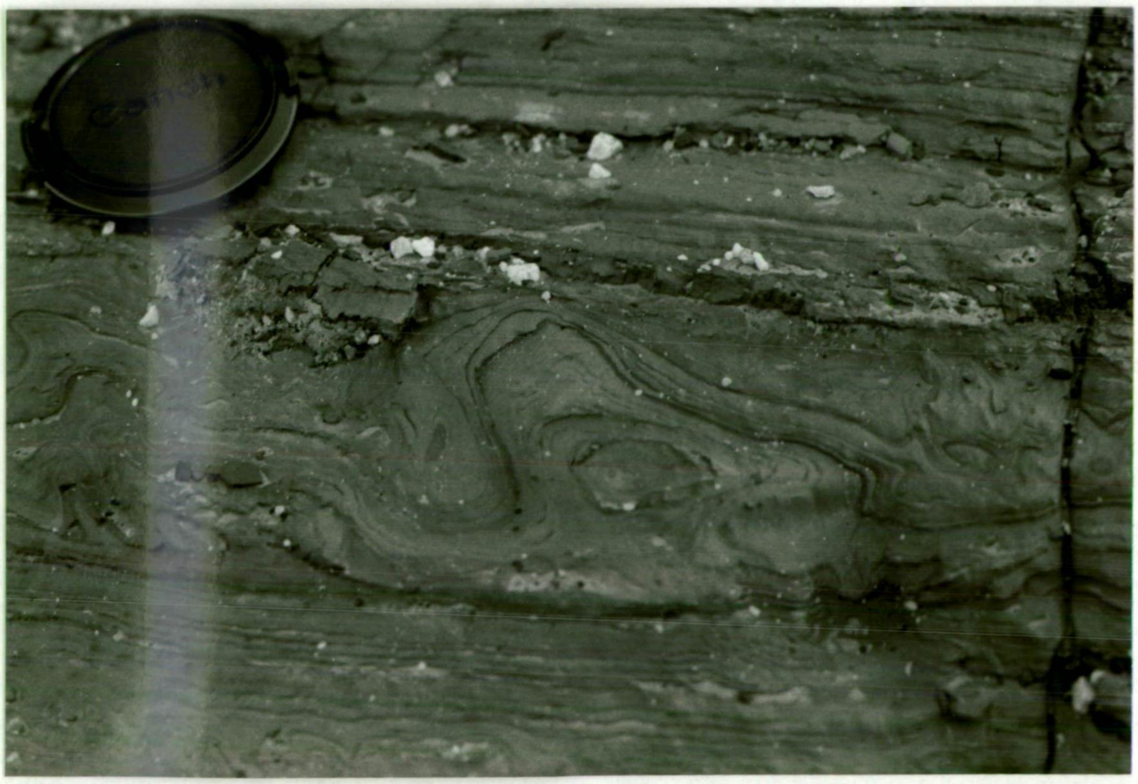


Fig. 6.10. Syndepositional convolute lamination of silts of the Thureau Formation type section.

(Fig. 6.6, fabric 7). This suggests that the disturbance may be due to coherent slumping associated with turbidity currents that flowed northward across the floor of the lake.

Numerous high-angle reverse faults in the silts are strongly clustered toward 279° and dip sub-vertically (Fig. 6.6, fabric 8). These faults were probably associated with other deformation structures observed from the section. On the southern side of the section a normal fault with a throw of 6.5 m is inferred from the position of intensely folded laminated silts near the bottom of the section (Fig. 6.4). A wedge-shaped block failure that is down faulted by 1.1 m and intrudes the underlying diamicton and sands also suggests a period of deformation. The structure bifurcates at a depth of 10 m and its margins are marked by humic acid and iron staining (Fig 6.4 and 6.11).

Unit 7.

Unconformably overlying the silts are light grey, poorly sorted, slope deposits derived from the Thureau Hills (Fig. 6.4). Although the contact with the underlying lake silts is certainly erosional, there is no indication of the period of time that separates their deposition.

The pebble fabric of the gravels, measured from the A/B dip direction of blade-shaped clasts according to the method of Rust (1972), shows a bimodal girdle pattern (Fig. 6.6, fabric 6). The particle size of the gravel is coarse and deficient in fines which reflects the siliceous source, the short distance, and high energy environment of transportation.

Interpretation.

The sequence records an ice-contact, deglacial environment. The sediments were probably deposited in a supraglacial position on melting ice. The reconstructed sequence of events in the deposition of the sediments is as follows. The terminus of the glacier became almost stagnant and deposition of a massive sediment flow or deformation of a melt-out till occurred. Then deposition of multiple sediment flows and sheet flows occurred within a depression that became filled with

Fig. 6.11. Wedge shaped failure and marginal chaotic deformation in Thureau Formation laminated silts.

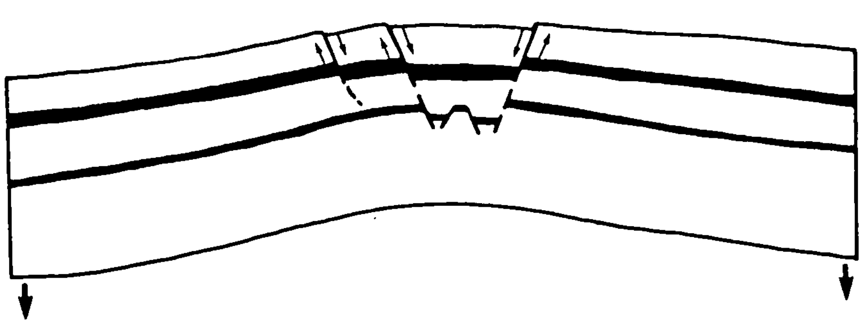


Fig. 6.12. Wedge shaped failure of type 1 experiment of Sandford (1959). The model simulates a broad upwarp, the equivalent of a loss of support by ice melting at the ends of the section.

water which increased in depth as the laminated sands accumulated. Subaqueous debris flows occurred within the developing kettle hole and were followed by the accumulation of laminated silts which settled out from small turbidity currents. The erosion and burial of the glacial sediments by paraglacial slope deposits occurred during and after the deformation of the entire section when the buried glacier ice melted.

The change from subaerial sediment flows to subaqueous deposition may represent sedimentation in a deltaic environment as the depth of water gradually increased.

Deformation of the sediments is very similar to Sandford's (1959) type 1 model experiments which have been adapted for explanation of ice-contact deformation by McDonald and Shilts (1975). The model (Fig. 6.12) shows the effect of gentle upwarping, which is the equivalent of removal of support at the ends of the section by melting. The melting caused the formation of a wedge-shaped graben. The central wedge failure in the Thureau Hills section is almost identical. Faults in the chaotically deformed laminated sediments on the southern end of the section suggest that the size of the buried ice blocks must have been large to have caused a throw of 6.5 m. However, such deformation is very localised and large parts of the section are relatively undeformed.

The Thureau Hills section.

Several other outcrops of the Thureau Formation occur in road cuts on the eastern side of the Thureau Hills. The section described here occurs at G.R. 882348 and is typical of these sediments.

Approximately 600 m south of the type section, up to 5.7 m of massive diamicton similar to Unit 1 (above) underlie up to 6 m of coarse slope deposits (Fig. 6.13).

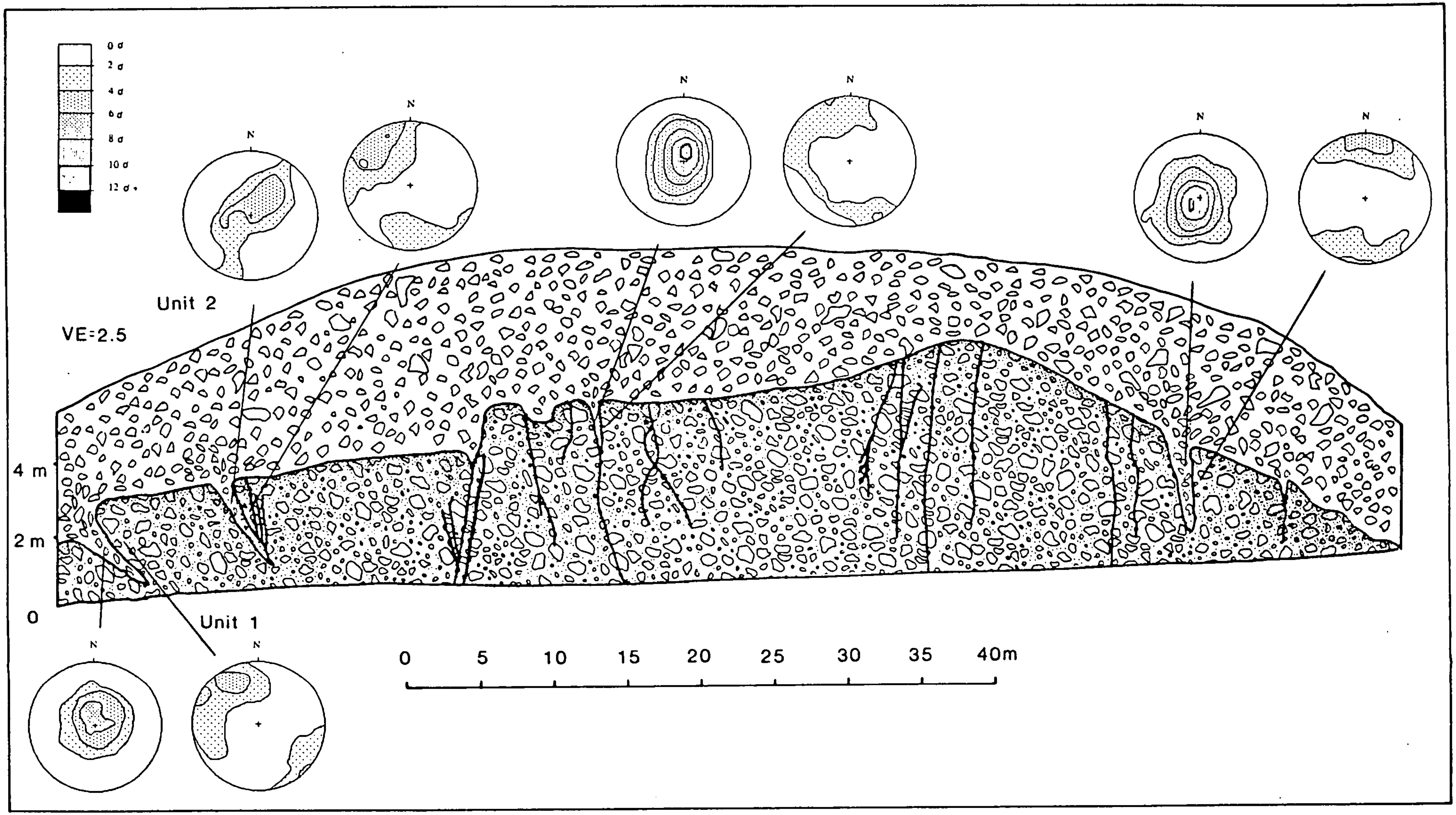


Fig. 6.13. Section through massive Thureau Formation tills overlain by slope deposits, 150 m south of the Thureau Formation type section. The upper part of the till is stained by humic acids and penetrated by wedge-shaped clastic dykes.

Fig. 6.14. Wedge shaped infill structures in Thrueau Formation tills. All dykes have humic acid and iron stained boundaries and prominent vertical clast alignment.

Unit 1.

Rudimentary, discontinuous bedding that dips south at 13° is the only structure in the coarse matrix-supported diamicton. Boulders up to 2.1 m in diameter have a mean weathering rind value of 29.8 mm and a standard deviation of 7.8 mm. Although the sediments are clearly the same age as the other Thureau Hills sediments, there is a large difference in the thickness of weathering rinds on Jurassic dolerite clasts. This suggests that other factors that operated locally may have influenced the weathering rate over long periods of time as was described in Chapter 5.

The pebble fabric of the diamicton is consistently weak and has a direction of maximum clustering toward 310° (Fig. 6.13).

Several wedge-shaped sedimentary dykes and narrow zones of humus-iron-manganese staining occur in the diamicton close to its contact with the overlying sediment (Fig. 6.14). Because humic acids have stained both the top of the diamicton and the sides of the cracks, it is unlikely to represent a truncated palaeosol. It probably represents leaching of humic acids through the porous slope deposits and their accumulation at the surface of the largely impermeable diamicton. Narrow zones of humic staining also appear to have penetrated down cracks in the diamicton up to 6 m deep. Over 80 sedimentary dykes have been observed in the King Valley and have been described in detail in Appendix 2. At Thureau Hills, the lithology, pebble fabric and particle size distributions distinguish the dyke fills from the host sediment (Fig. 6.14). The dykes have been filled from above by the overlying slope deposits. The fills have no erratic clasts of Jurassic dolerite and Permian sediments which are abundant in the host diamicton. The dyke-fills have much coarser particle size distributions than the host sediments (Fig. 3 in Appendix 2). This reflects the siliceous nature of the fill sediment, and the different transport mechanisms between the fill and host sediment. The pebble fabrics of the fills shows a pronounced near vertical orientation imparted during the fill phase of dyke formation (Figs. 6.13 and 6.14). Because rapid filling by slumping would give a chaotic fabric the near vertical fabric suggests that filling was slow.

The origin of these dykes is thought to be associated with the other syndepositional ice melt features and to have formed in a similar manner to boulder-filled cracks observed forming on supraglacial debris of Icelandic glaciers by Eyles (1979). The origin of these and other clastic dykes in the King Valley is discussed in Appendix 2.

Unit 2.

The diamicton is overlain by a coarse, massive slope deposit which is the source of the infills of the sedimentary dykes.

The slope deposit consists entirely of Siluro-Devonian sediments, Ordovician conglomerates and Cambrian volcanics from the adjacent Thureau Hills. It is essentially the same deposit that overlies the glacial sediments of the type section.

Sections on the southern end of the series of exposures along the Thureau Hills show clast-supported diamictons with crude horizontal bedding and better sorting. These sediments appear to be the proximal outwash gravels deposited downstream of the ice contact sediments in this section and the type section described above. Together the series of sections appear to record an ice terminal position.

Interpretation.

These sections and several others exposed at Thureau Hills consist mainly of massive tills that appear to have been disturbed by mass movement during and after deposition. They have almost certainly been deposited in a supraglacial position above melting ice. Although Boulton and Eyles (1979) and Eyles (1979) give detailed descriptions of sediment accumulation in supraglacial positions, there are few similarities between their descriptions of supraglacial sediments and those observed at Thureau Hills. Nevertheless in ice-contact situations local sedimentary environments are both complex and variable, and can give rise to different forms and sediments at different localities.

The Regency sections.

Several sections in the lower King Valley between the Governor and King rivers show major unconformities between sediments of the Thureau Formation and outwash gravels of the subsequent King and Blackwood advances. There are two sections where the features are well exposed in a quarry at G.R. 876312 (Fig. 6.15 and 6.16). In the first section a lens of organic debris rests conformably on Thureau Formation outwash gravel. It is unconformably overlain by Blackwood Formation outwash gravels that appear to be reworked from older till. Pollen analysis of the organic sediments shows an interglacial floral assemblage, which is the Regency Interglacial (section 4.4). The second section shows a similar unconformity between massive till of the Thureau Formation and Blackwood Formation outwash gravels.

Description of section 1.

Unit 1.

The lowermost sediments of this section consist of highly weathered, massive, moderately well sorted outwash gravel. Pebble fabric of the dip direction of the A/B planes of disc-shaped clasts shows a weak preferred orientation toward 341°, suggesting deposition by a stream flowing approximately from north to south (Fig. 6.15). All Jurassic dolerite clasts in this deposit are completely weathered and reduced to a dull yellow clay. On the western end of the section a 30 cm-thick dyke of gravel strongly cemented by humic acid precipitate penetrates the host gravel. The dyke can be traced from one wall across the floor into the opposite wall of the excavation. The truncation of this dyke at the top of the gravel demonstrates the presence of a land surface that has now been completely eroded.

Clasts from the gravel of unit 1 (Fig. 6.15) are mixed with the overlying interglacial organic deposit and suggest that the contact is conformable. This interpretation is supported by the pollen analysis which suggests succession from a boggy environment with standing water to temperate

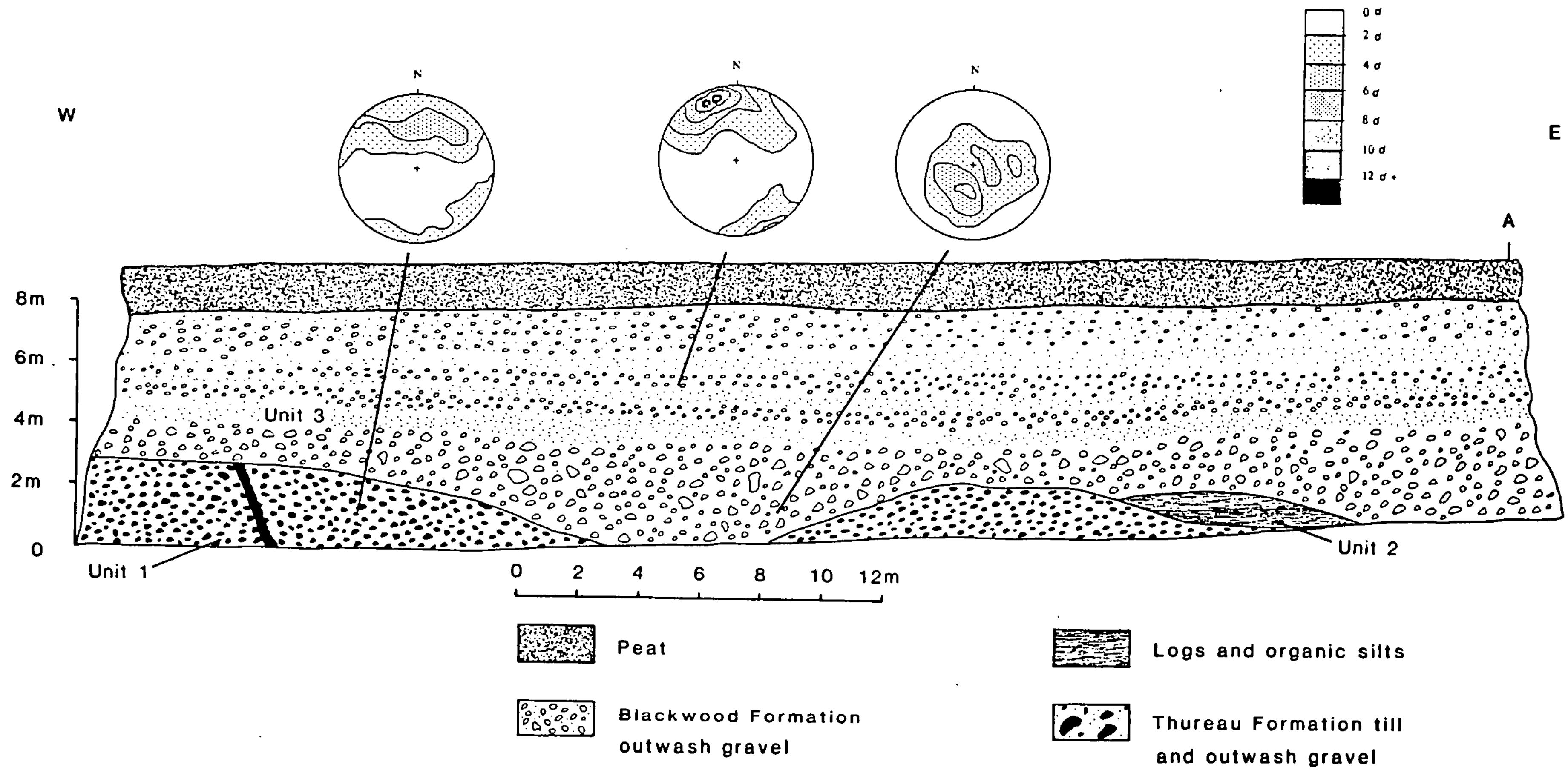


Fig. 6.15. Section through Blackwood Formation outwash gravel overlying a lens of Regency Formation organic sediment that rests on highly weathered Thureau Formation outwash gravel. The Thureau Formation outwash gravel contains a narrow dyke of humic acid precipitate

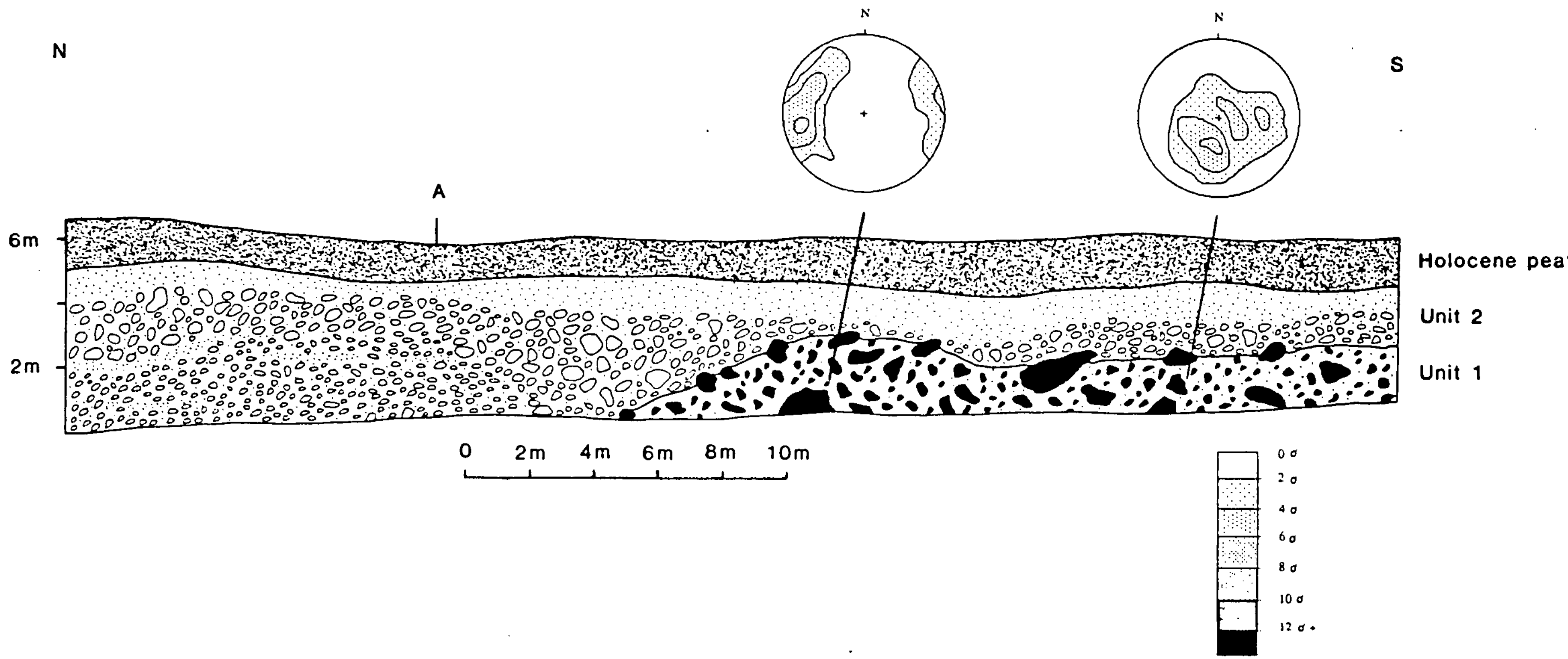


Fig. 6.16. Section through Blackwood Formation outwash gravel resting on highly weathered Thureau Formation till. This section intersects the section drawn in Fig. 6.15 at 90° at the point A marked on Fig. 6.15.

A

B

Fig. 6.17. The Regency sections. A, Blackwood Formation outwash gravel overlying a lens of Regency Formation organic sediments. B, Blackwood Formation outwash gravel resting on highly weathered Thureau Formation till.

rainforest. The initial wet environment probably developed on a low-lying, uneven, recently deglaciated surface.

Unit 2.

The organic deposit consists mainly of a combination of *in situ* humified peat matter overlain by drifted wood and leaves. The organic content is very high and ranges from 70 to 95% loss on combustion. The size of the organic lens appears to have been greatly reduced by erosion. The contact with the overlying gravel is sharp, dips west at up to 10° and is eroded.

Unit 3.

Part of the overlying gravel is, poorly sorted gravel with a maximum particle size of 500 mm and a well sorted with a maximum particle size of 80 mm. The two gravels are separated by a 0.3 m-thick bed of massive medium sand (Fig. 6.17a). The lower, coarser gravel is massive and consists of large blocks of a wide range of rock types that are supported by a matrix of coarse sand. Diffuse iron staining has altered the colour of the gravel to a mottled yellow colour and hard iron cemented layers have also been formed.

The overlying well sorted sand is massive, free flowing, and contains occasional isolated pebbles and pebble bands. The sand is a light grey colour and is unaffected by the iron staining, as is the overlying gravel. The contact with the overlying gravel is sharp and appears to have been scoured by meltwater streams.

The overlying 1.8 m of gravel consists of well sorted pebble gravel with a maximum particle size of 70 mm, interbedded with poorly sorted coarse sand and pebbly sand. The pebble fabric of the gravel shows a bimodal pattern with the direction of maximum clustering toward 339°, which suggests that crude horizontal bedding was formed by a current flowing from northwest to southeast. Weathering rinds on the Jurassic dolerite clasts have a mean thickness of 1.9 mm and a standard deviation of 0.5 mm, which is considerably less than many other deposits of this age.

The anomalies in weathering of deposits in this part of the valley have been discussed in Chapter 5.

A thin blanket of peat up to 0.8 m thick overlies the gravels and forms a continuous cover on the surrounding low-lying plain.

Description of section 2.

The second Regency section (Fig. 6.16) is oriented north to south and is intersected at right angles by the eastern end of the first Regency section at the position marked on Figure 6.15.

Unit 1.

The basal sediments of this section include a massive, highly weathered till that rests on a light grey massive clay. This clay appears to be part of the weathered Ordovician limestone that underlies most sediments in this area. Erratic boulders of Jurassic dolerite, Ordovician conglomerate and Permian sediments up to 1.5 m in diameter are common in the till (Fig 6.17b). Weathering rinds on Jurassic dolerite clasts have a mean thickness of 75.5 mm and a standard deviation of 14.5 mm. The pebble fabric of the diamicton is weak and is unrelated to the reconstructed ice flow direction (Fig. 6.16). The second fabric shows that the clasts have a very steep dip, which is not characteristic of any known origin for glacial deposits. It may be due to localised deformation.

The contact with the overlying gravels is uneven, sharp and eroded. Where large weathered dolerite boulders protrude through the till their weathering crust has been removed by scour associated with the currents that eroded the till and redeposited it as gravel.

Unit 2.

The gravel is coarse and poorly sorted with particle sizes of up to 800 mm in diameter. It contains occasional lenses of sorted sand and resembles the upper gravel in the Regency 1 section. The gravel grades upward into poorly sorted sand which is overlain by up to 0.6 m of fibrous peat. The surface of the deposit is a relatively flat terrace marked by several sinuous channels that have a braided pattern. The surface of the terrace is marked by several clusters of large erratic boulders up to 2.5 m in diameter which appear to be lags from the erosion of Thureau Formation tills.

Interpretation.

Comparison of the two Regency sections shows that the Thureau Formation tills in the second section are 1 to 2.5 m above the Thureau Formation outwash gravels in the first section. This suggests that the till protrudes through outwash gravel. The sediments may be part of a supraglacial sediment association with massive tills protruding through outwash gravels or they may be basal tills that protrude through outwash gravels. The pebble fabric suggests that the till has been significantly modified by secondary processes after its release from the ice as the orientation of clasts bears no relationship to the direction of ice movement (Fig. 6.16). Comparison of the fabrics with those for secondary processes observed by Lawson (1979a) suggests the first fabric is more typical for sediment flows and the second may be some kind of dump deposit that has accumulated as ice slope colluvium by falling or sliding off an ice surface. The former alternative of deposition in a supraglacial position is considered the more likely of the two depositional environments.

The differences in weathering rind thickness on the Jurassic dolerite clasts in the Thureau Formation sediments of the lower King Valley and at Thureau Hills suggest that a significant period of time separated their deposition (Fig. 6.18). Because the sediments exposed at Thureau Hills record an ice terminal position they may represent a different ice advance to the sediments in the lower King Valley. However,the variability in weathering rind thickness in the Thureau Hills sediments (Fig 6.18) suggests that small scale site factors may have had a significant effect on weathering rates (Chapter 3). If so, inferences about the relative age of deposits in the upper and lower King Valley are likely to be of dubious quality. Unfortunately, because all the depositional

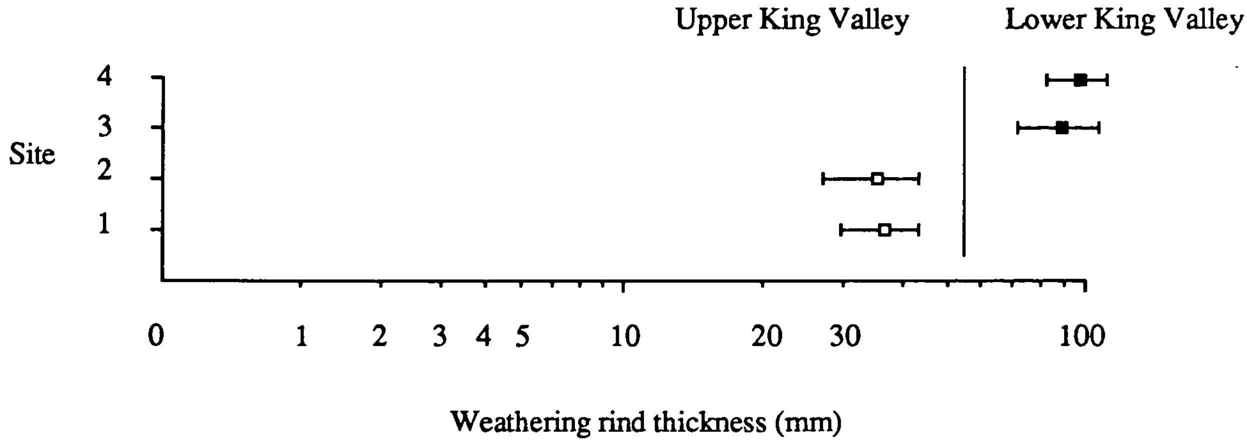


Figure 6.18 Differences in the thickness of weathering rinds between Thureau Formation sediments in the upper and lower King Valley. Each point represents the mean of 50 measurements and the bar represents ± 1 standard deviation

morphology is eroded in both areas there is no way to demonstrate that the deposits may belong to separate ice advances. Future research may clarify this situation.

The organic sediments that lie above the Thureau Formation sediments in the section record the only pollen evidence of interglacial warming between the Thureau Formation and younger sediments. The geometry of the organic lens suggests that it was formally of greater extent and has been eroded by sediment fluxes associated with the succeeding glacial advances.

The interpretation of the overlying sediments as Blackwood Formation outwash gravel is based on the recognition of the aggradation surface being inset into adjacent, topographically higher King Formation outwash gravels. An interpretative cross-section through the valley at this point shows the relationship between the Blackwood, King, and Thureau Formations and the Ordovician limestone (Fig. 6.19).

Newall Creek sections.

At Newall Creek a series of outwash deposits have been deposited near the King River on the western side of the West Coast Range at G.R. 784322. The narrow steep-sided King River Gorge through the West Coast Range acted like a large flume during the melt of glaciers in the King Valley. Terraces downstream of the mouth of the gorge are evidence of the large amounts of sediment that passed through the gorge. Although the age of the deposits has not been accurately determined they are the most weathered outwash gravels in the study area. Although the laminated clays in the upper part of the section have been analysed there is no clear direction of magnetic polarity (M. Pollington pers. comm. 1987). However, the degree of weathering of the sediments suggests they may correlate with the Thureau Formation tills on the eastern side of the West Coast Range.

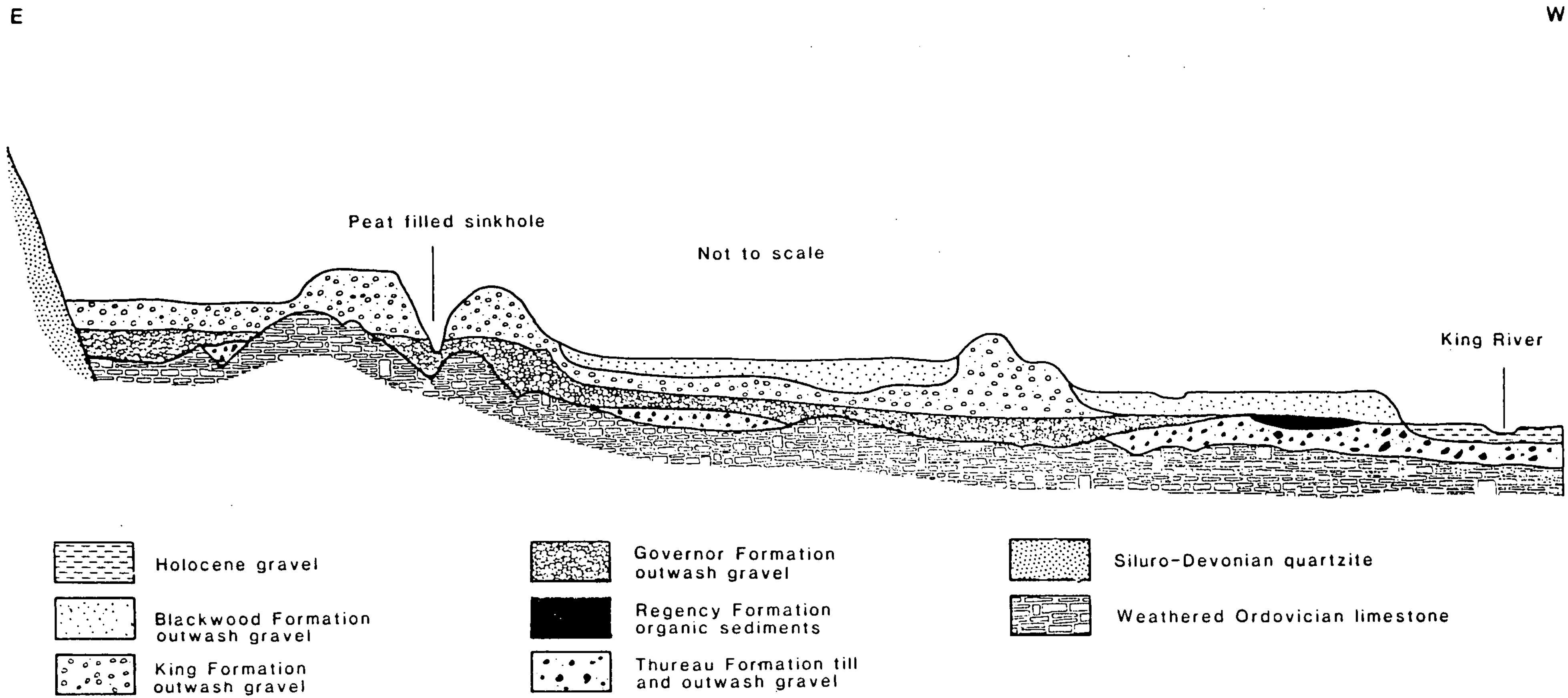


Fig. 6.19. Reconstructed relationship and geometry of the contacts between the King, Blackwood, Regency and Thureau formations in the lower King Valley.

Description.

A section on the left bank of the King River shows a series of laminated silts sands and gravels (Fig. 6.20). The bottom unit consists of coarse, poorly sorted, massive iron-stained gravel with a mean particle size around 60 mm. The gravels are highly weathered with rinds on Jurassic dolerite up to 28 mm thick. The contact with the overlying sediments is gradational from gravel to silty gravel, clayey gravel to clay.

The white silty clay overlying the gravel is laminated and very compact. It contains numerous small dispersed granules and coarse sand particles. The clay grades laterally into silty sand on the western side of the section over a distance of 3 m and forms part of the same elliptical geometry as the overlying sands and silts (Fig. 6.20). The clay grades upward from massive clay to interbedded white clay and silt, to light brown sandy silts (Fig 6.21). The clay has a number of high-angle reverse faults (Fig. 6.20 and 6.21). On the increase in the throw of the faults, and the brecciation of the clay bed and minor boudinage structures suggest that the intensity of deformation increased from west to east. The contact with the overlying sediments is sharp.

The uppermost sediments on the western side of the section consist of a series of silts and sands with an elliptical geometry. The top 1.5 m of these are black with humus and iron staining and are weakly iron cemented. On the eastern side of the section 2 m of massive, medium sand lies on laminated sand into which the reverse faults extend.

One hundred and fifty metres to the south and apparently underlying the section described above, an 18 m section shows a sequence of massive, highly weathered, coarse outwash gravels (Fig. 6.22). The lithology of the gravels is mixed and shows a significant contribution of erratic Jurassic dolerite and Permian sediments. The gravels have a distinct textural variability with a thick bed of coarse bouldery gravel overlain and underlain by weakly bedded gravelly sand.

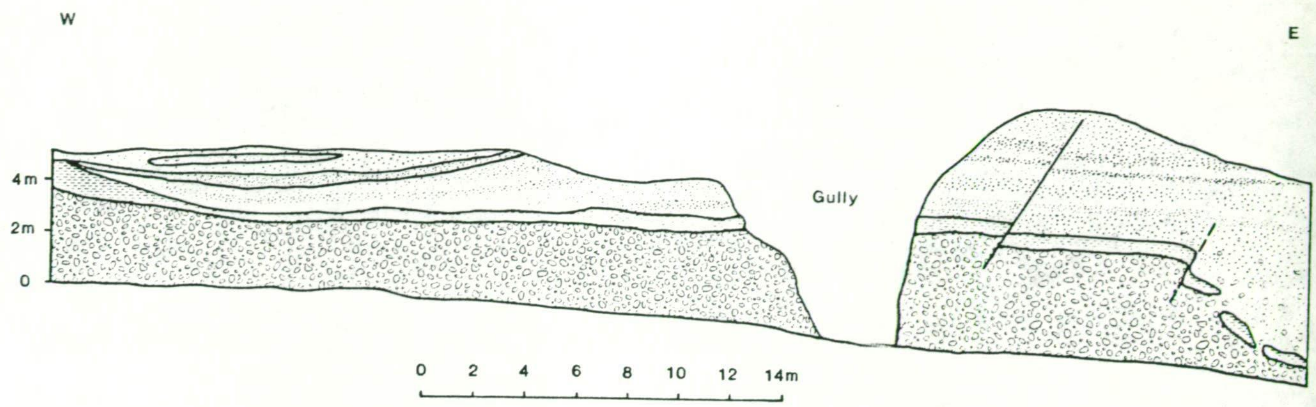


Fig. 6.20. Thureau Formation sediments exposed at Newall Creek. Coarse outwash sands are overlain by faulted, laminated clay, mud and sandy silt.

Fig. 6.21. High angle reverse fault in sediments exposed at Newall Creek.

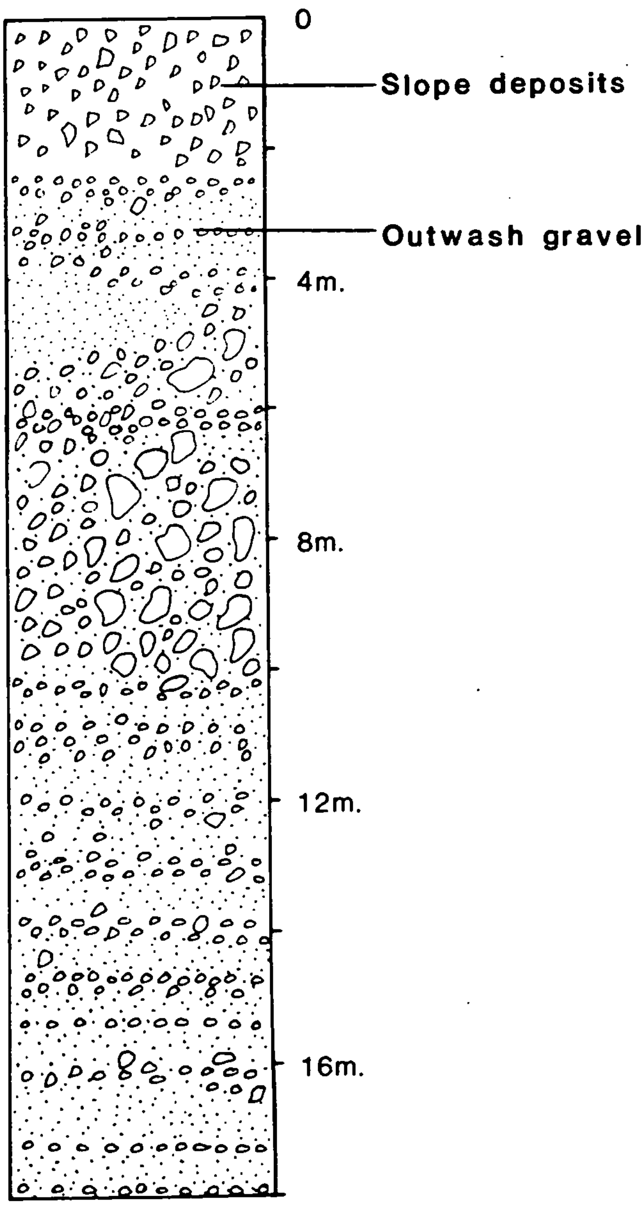


Fig. 6.22. Sequence of sediments that underlie the section shown by Fig. 6.20.

Interpretation.

The lithology of the gravels shows a large proportion of erratic clasts that have come from either the King or Queen River valleys. The extent and sources of ice in the Queen Valley is unknown. An unusual pattern in the lithology of these gravels is the presence of Jurassic dolerite and Ordovician conglomerate boulders over 350 mm in diameter. These boulders are over 7 km from the closest known ice source. Their presence is probably due to the confining conditions imposed by steep rock walls of King Gorge. Because the river reach had a high competence, it was capable of transporting large volumes and sizes of sediment. The outwash gravels become increasingly finer downstream from the gorge exit to the river mouth in Macquarie Harbour.

The silts and clays in the terrace-top exposure may have been formed as an outwash terrace from the King River aggraded and formed a small pool. The origin of the post depositional deformation of these sediments is unclear. Glaciers are not known to have flowed this far south in the Queen Valley, so the faulting is not likely to have developed from deposition on melting ice. Because the section is within 20 m of the edge of the terrace which lies 35 m above the King River, it is possible that the faulting has developed from slumping caused by removal of support by river erosion. Figure 6.21 shows that the fault plane is open and appears to be under tension, which supports the interpretation of contemporary slumping toward the terrace edge.

Gormanston Football Field section.

Extensive highly weathered Thureau Formation moraines occur in the upper part of the Linda Valley. Most of the sediments consist of chaotically deformed ice-rafted diamictons and sediment flows interbedded with laminated lake silts. The section described here from G.R. 836412 is representative of these sections.

Description.

The section at Gormanston Football Field consists of dark grey laminated lake silts interbedded with ice-rafted diamictons and flow tills that rest on highly weathered massive till (Fig. 6.23). The silts have a reversed detrital remanent magnetisation (Barbetti and Colhoun 1988 in press and M. Pollington pers. comm. 1987) and are believed to lie in the later part of the Matuyama Chron.

Unit 1.

At the base of the section, highly weathered Jurassic dolerite boulders up to 1.8 m in diameter rest in a silty matrix. The diamicton overlies Cambrian volcanic rock detritus which is believed to have resulted from ice fracturing.

Unit 2.

The massive till is overlain by 1.5 m of grey laminated silts with numerous dropstones (Fig. 6.23 and 6.24a). Within the silts are discrete beds of ice rafted pebbles that appear to record episodes of ice rafting of boulders. The fabric of the ice rafted pebbles is weak with many clasts having a steep dip (Fig. 6.23).

On the eastern side of the section thick gravel lenses occur in the silt. These isolated lenses are associated with intense deformation of the silt which is folded, faulted and brecciated (Fig. 6.24). The pebble fabric of the clasts is weak with the direction of maximum clustering toward 100°. The intense marginal deformation around the lenses suggest they formed as subaqueous sediment flows.

Unit 3.

The silts are overlain by laminated brown silty sand. The contact with the underlying silt appears to be scoured.

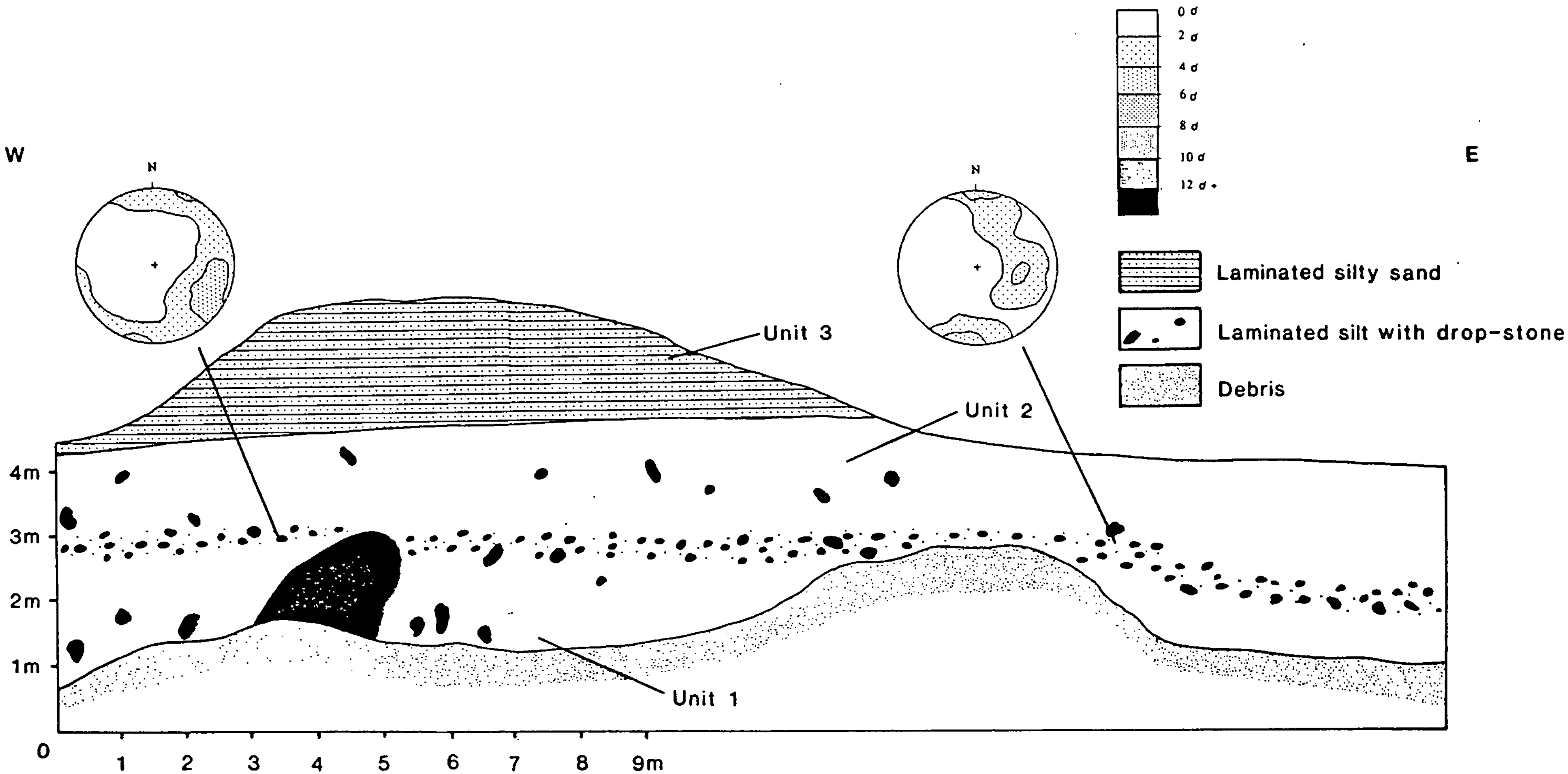


Fig. 6.23. Thureau Formation sediments exposed at Gormanston Football Field. Laminated sandy silts overlie laminated mud containing sediment flows and ice rafted diamictons which rest on massive till.

A

B

C

Fig. 6.24. Thureau Formation sediments at Gormanston Football Field. A, Ice-rafted pebbles in gently warped laminated mud. B, Ice rafted till overlain by a massive debris flow with intraformational blocks of laminated mud, and overlain by faulted laminated mud. C, Detail of deformation at the base of the sediment flow shown in B.

Interpretation.

The section records subaqueous deposition in a proglacial or supraglacial lake that was subject to ice rafting and occasional subaqueous sediment flows. The scoured surface and sudden change in sediment size in unit 3 may be related to the retreat of the ice margin and the lake subsequently becoming shallower.

6.5 The Traveller Formation.

The Traveller Formation is known from remnants of outwash gravels that rest topographically above the aggradation level of the Fish Formation. Sediments of this formation are not very extensive in the King Valley and are for the most part indistinguishable from sediments of the Fish Formation. However, at Baxter Rivulet they are separated from Fish Formation sediments by organic sediments of the Baxter Interstadial. Because the sediments are derived from Mt. Jukes their extent is limited to the lower King Valley but outwash gravels do extend into the Andrew River. The deposits have not been mapped beyond the King River catchment.

The extent of the ice advance associated with the sediments is not known, but it is thought that it may have breached the Baxter Rivulet catchment and entered the Governor River. Lack of exposure in the Governor River make this impossible to confirm.

The type section.

A section on the right bank of Baxter Rivulet at G.R. 875300 shows Traveller Formation outwash gravel resting beneath organic silty sands that record an interstadial flora. The same section shows sediments from the Fish Formation overlain by outwash gravel from the Governor and King formations (Fig. 6.25).

Description.

Unit 1.

The basal sediment of the section is highly weathered till that crops out as narrow dykes in the surface of the weathered Ordovician limestone (Fig. 6.25). Boulders of Jurassic dolerite up to 150 mm are completely weathered and the till is thought to correlate with the Thureau Formation which has been observed to occur as fillings of solution tunnels elsewhere in the valley (Fig.

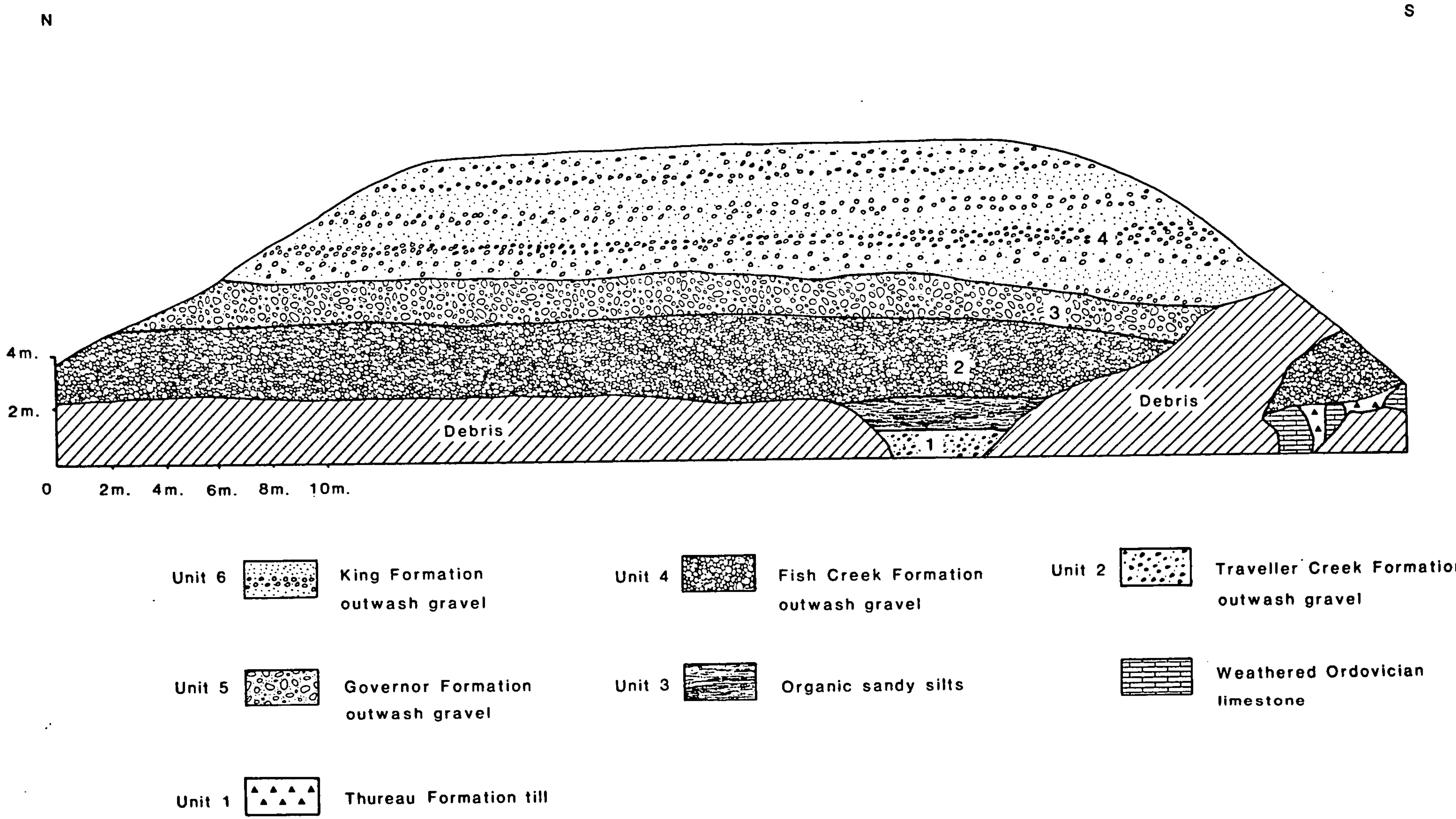


Fig. 6.25. The Baxter Rivulet section. A major unconformity is present between units 5 and 6.

6.59). The contact between the till and Traveller Formation gravels was not observed at this section.

Unit 2.

Gravel of the Traveller Formation is seen in the middle of the section in a shallow excavation. The gravel is well sorted and rounded. The pebble fabric and lithology of the gravel indicates that it was deposited by a stream flowing from Mt. Jukes (Fig. 6.26-2 and Fig. 6.27-1).

Unit 3.

Overlying the outwash gravel is 1.2 m of organic sandy silt that records an interstadial palynoflora of the Baxter Interstadial (see Chapter 4). The silt is horizontally bedded and contains pieces of drifted wood up to 60 mm long. The organic content of the silt, measured as loss on combustion in an oven at a temperature of 700° C for 4 hrs ranges from 1.6 to 10.7% (Fig. 6.28). The increase in the organic content of the sediments at 50 cm corresponds to the boundary of pollen zone BR 1 which is characterised by the maximum development of herbs (Fig. 4.3). The contact between the silt and overlying gravel is sharp, but the silt does grade into coarse sand in the uppermost 15 cm.

Unit 4.

The Fish Formation outwash gravel overlies the organic sediments and consists of up to 1.7 m of massive, well sorted and rounded clast supported gravel with a maximum particle size of 60 cm. The lithology and pebble fabric of the gravel is very similar to that of the Traveller Formation and is consistent with deposition by a braided river that flowed from the West Coast Range. The contact between this gravel and the overlying bouldery gravel is sharp and appears to be scoured.

Unit 5.

The 1.6 m of bouldery gravel that overlies the Fish Formation outwash gravel is part of the Governor Formation. It consists of coarse, poorly sorted gravel with boulders up to 700 mm in diameter. The lithology suggests the gravel is a deposit from the King Glacier (Fig. 6.26-3). The

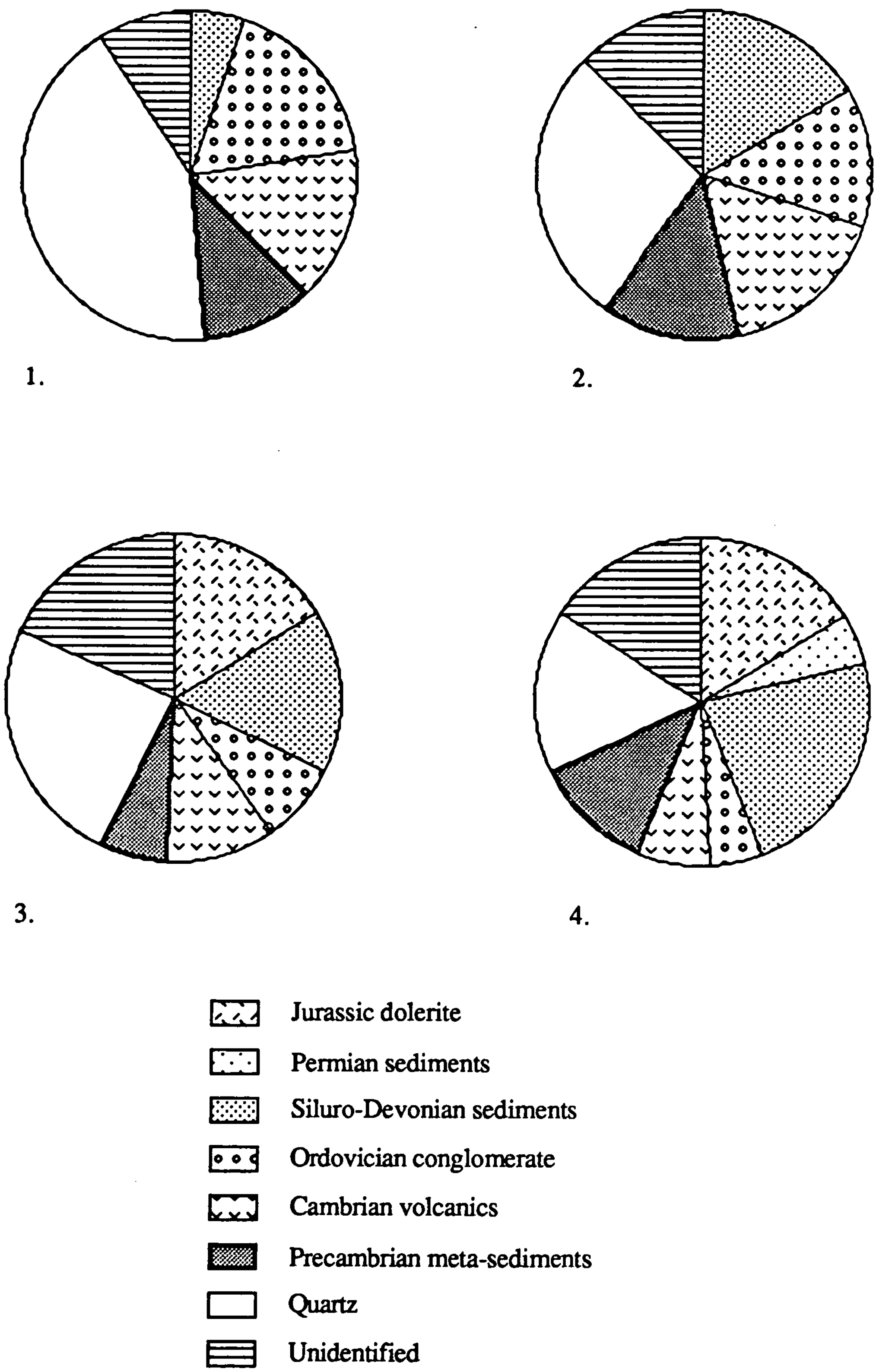


Fig. 6.26. Lithology of the sediments exposed at Baxter Rivulet. 1, the Traveller Formation. 2, the Fish Formation. 3, the Governor Formation. 4, the King Formation. See Fig. 6.25 for sample locations.

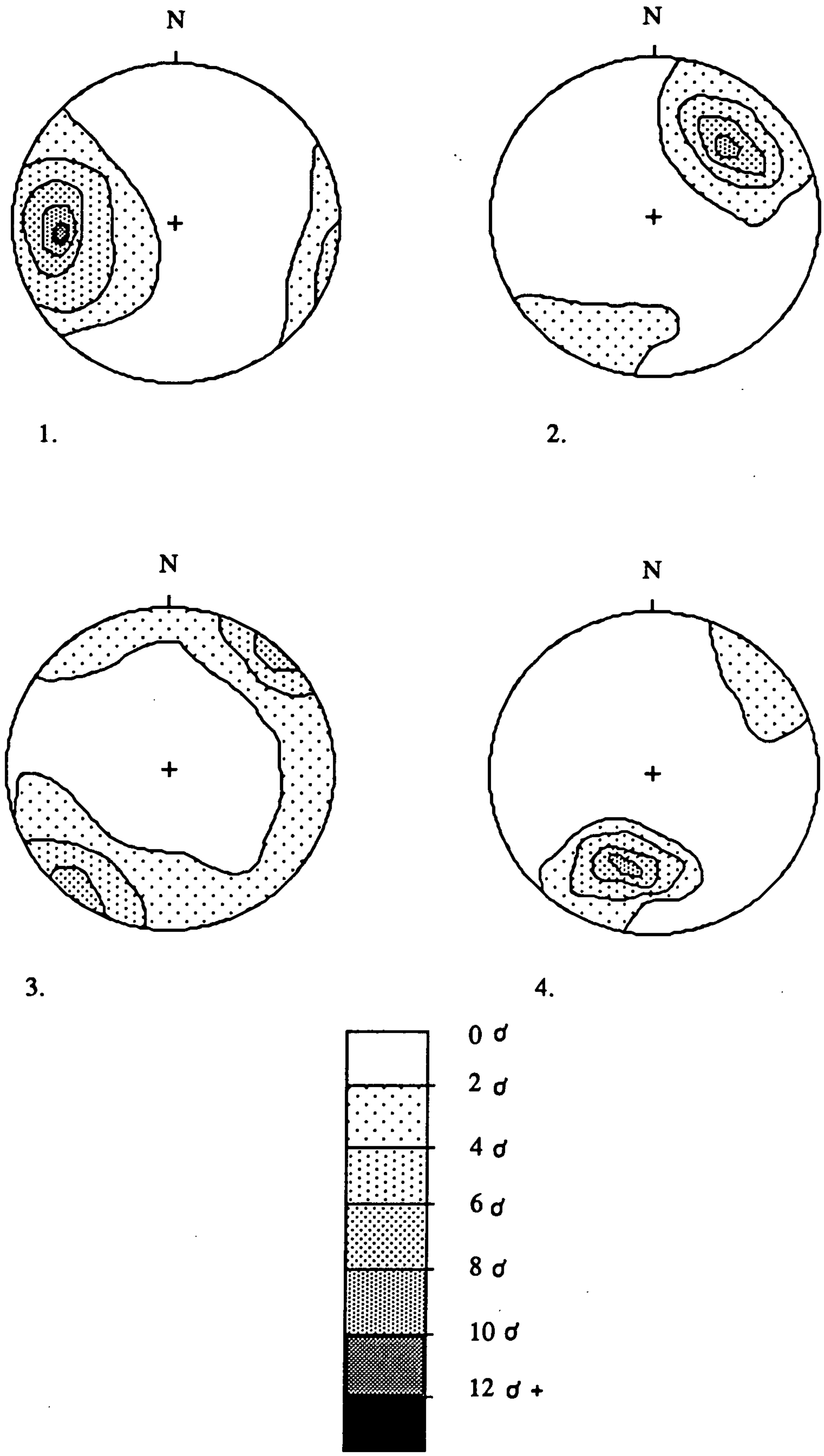


Fig. 6.27. Pebble fabric of the sediments exposed at Baxter Rivulet. 1, the Traveller Formation. 2, the Fish Formation. 3, the Governor Formation. 4, the King Formation. See Fig. 6.25 for sample locations.

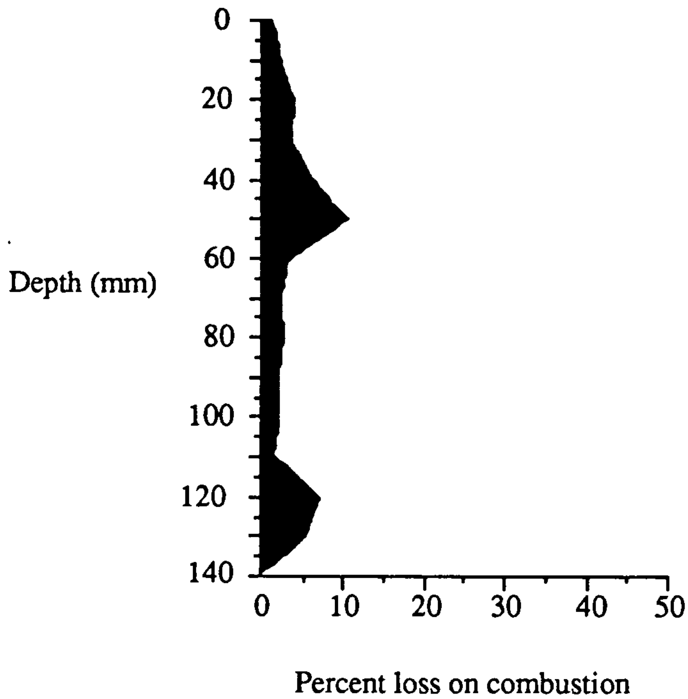


Fig. 6.28. Loss on combustion of the Baxter Formation sediments. The 10 cm sample is the same level as the 0 cm pollen sample in Fig. 4.3.

pebble fabric suggests the gravel has been derived and redeposited from a till (Fig. 6.27-3). The contact between these gravels and the overlying outwash sediments of the King Formation is sharp, and from the protrusion of larger boulders through the contact it appears to have been scoured. The contact between the Fish and Governor Formation sediments can be traced southward where it manifests itself as a lithological boundary on the surface of the terrace. This boundary, which can be traced across the floor of the King Valley, is the southern limit of erratic dolerite and Permian clasts from the King Glacier .

The geometry of the contact between the two sediments appears to be due to the burial of the Fish Formation outwash surface by the larger sediment flux from a contemporaneous outwash fan of the King Glacier that formed during the Governor advance (Fig. 6.29).

Unit 6.

The King Formation outwash sediments consist of an upward-fining sequence of clast supported gravels, interbedded with coarse sand and gravelly sand. The lithology of the gravel is very similar to that of the underlying Governor Formation sediments except for a slight decrease in erratic lithologies toward the top of the section (Fig. 6.26-4). The pebble fabric of the gravel suggests deposition from a stream flowing from west to east, which is consistent with the slope of the outwash terrace and the position of the glacier terminus during the King advance (Fig. 6.27-4).

Interpretation.

The lithology and pebble fabric results show that the section at Baxter Rivulet records the superposition of sediment bodies from different source areas (Fig. 6.29). The evidence demonstrates that the outwash deposits from the King and Jukes glacier systems overlapped. However, it is not known whether the glacier fronts were ever in contact.

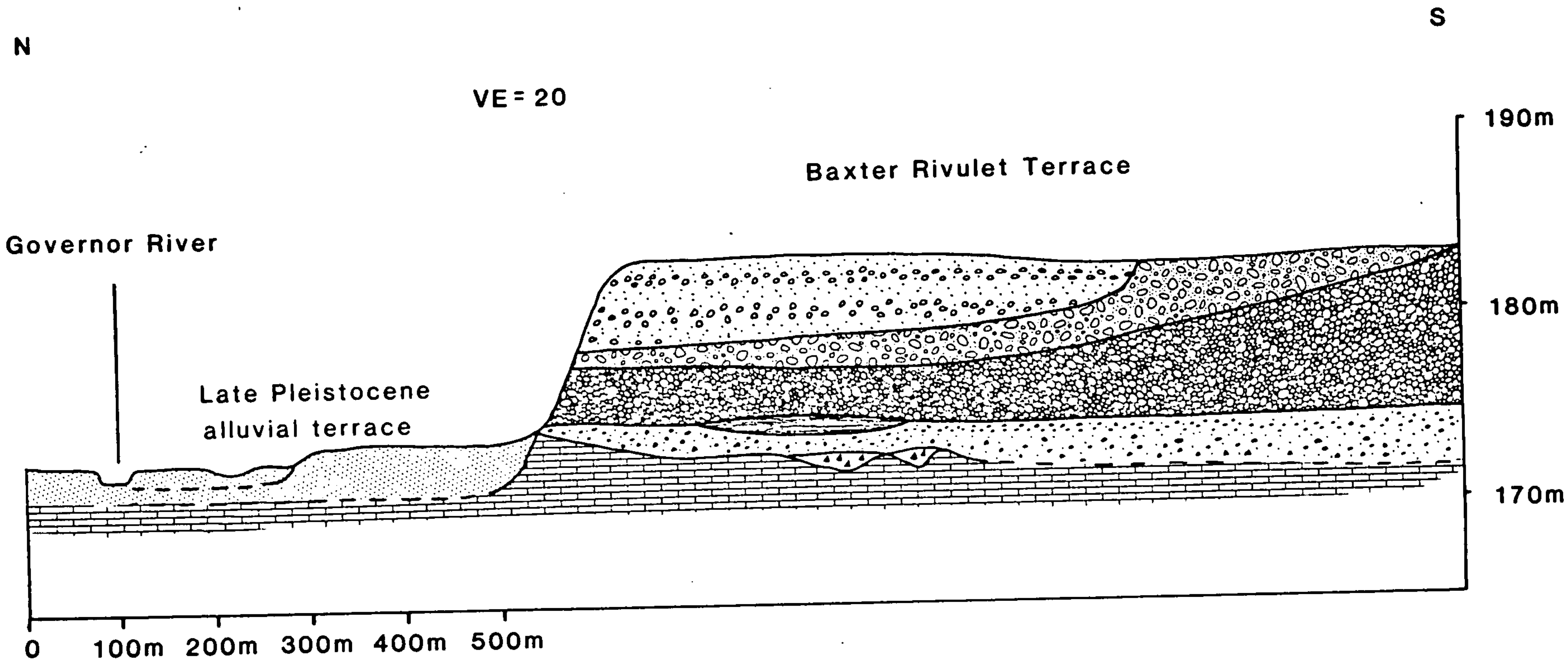


Fig. 6.29. Stratigraphic relationships in the Baxter Rivulet terrace. Symbols are the same as those used in Fig. 6.25.

The Fish and Governor Formations represent ice advances from different sources viz. Mt. Jukes and the Tyndall Plateau. Although the events are of slightly different age they have the same cause, a climatic deterioration leading to an ice advance. The degree of time lag in the response to climatic change depends on the nature of the geologic processes involved (Watson and Wright 1980). The lag in deposition of the Governor Formation over the Fish Formation is due to the different sizes of the King and Mt. Jukes glaciers. The area covered by the King Glacier was more than 25 times that of the Mt. Jukes Glacier during this advance. The response time for a given precipitation input to lead to an advance would therefore have been much shorter for the Mt. Jukes Glacier than for the King Glacier.

The sediments of the Fish and Governor formations were deposited at different times and have different lithological characteristics. Because the deposits were formed during the same general phase of glacial advance there are problems in stratigraphic classification and nomenclature. Stratigraphic practice in mapping Quaternary deposits is to classify these two formations as part of the same lithostratigraphic unit even though they were not deposited at the same time. Suggate (1965a) dealt with a similar problem in the Grey Valley in New Zealand by defining sediments from the same advance of different glaciers as one formation where their aggradation surfaces can be confidently correlated with each other. However, here the formations have been classified as separate lithostratigraphic units because they have different lithologies and relate to geographically distinct ice sources.

The most extensive glacial sediments in the King River Valley are those of the middle Pleistocene Comstock Glaciation (Kiernan 1983; Colhoun 1985a). In Tasmania it has long been suspected that both the Comstock and Linda glaciations represent more than two glacial stages (Kiernan 1983a). The evidence to support this suspicion has not been forthcoming. The section at Baxter Rivulet is direct evidence that sediments previously regarded as of Comstock Glaciation age contain a major unconformity that indicates a gap between deposition of the Governor and King formations in excess of 100 ka that probably represents interglacial warming. This is discussed in section 8.1.

The organic deposits at Baxter Rivulet are regarded as interstadial because they lie between two glacial outwash gravels and record an open forest environment with a significant contribution of subalpine plants. The Baxter Interstadial is defined by lithostratigraphy and biostratigraphy, whereas other interstadials in Tasmania are zones in pollen diagrams defined solely by changes in floral characteristics.

The section at Baxter Rivulet is important in the glacial stratigraphy of western Tasmania because it is evidence for a glaciation between the Linda and Henty Glaciations, it is the only site in Tasmania that shows clear evidence of the interaction of sediment bodies of two different glaciers, it shows that there has been geographically differentiated responses to changes in snow accumulation and climate change, it is the only site in Tasmania where interstadial flora is preserved between glacigenic sediments, and it is one of the few sites to record recognisable unconformities in Quaternary glacial sediments.

The Traveller Creek section.

A section in the Traveller Formation occurs in a small quarry east of Mt. Jukes at G.R. 862289. The section consists of massive, bouldery, proximal outwash gravels resting on weathered Devonian quartzite (Fig. 6.30). The gravel was deposited at the apex of the Traveller Creek outwash fan, probably during the retreat of the glacier into the Mt. Jukes cirque.

Description.

The gravel is stratified with imbricated pebble gravels overlying bouldery gravels that may be a lag from an eroded till (Fig. 6.30). Particle sizes up to 700 mm in diameter suggest that the gravel was deposited close to an ice front, or that it was glacially transported almost to the site and then reworked by outwash streams. The lithology of the gravels (Fig. 6.31) shows they were entirely

Fig. 6.30. Traveller Formation outwash gravel resting on Siluro-Devonian quartzite.

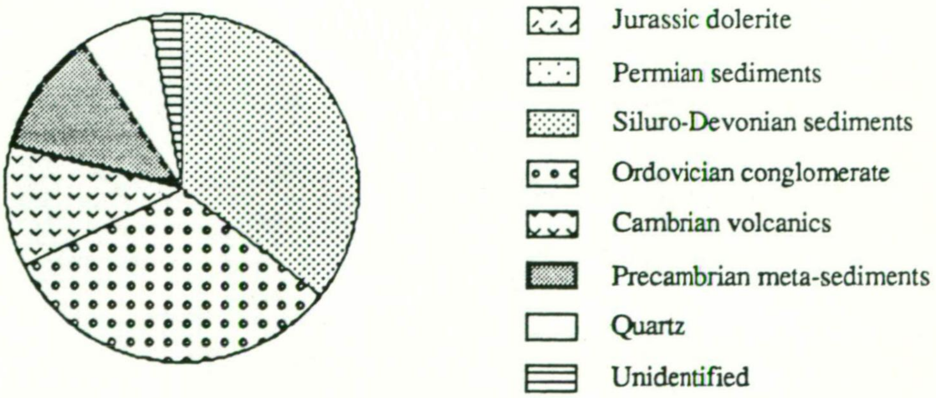


Fig. 6.31. Lithology of the Traveller Formation sediments shown in Figure 6.25.

derived from the West Coast Range.

Iron pans up to 1cm-thick have formed in the pebble gravels and below the pans the sediments are iron stained. Humic acids have accumulated above the iron pans, and at the boundary between the sediments and the rockhead. One kilometre east of the section a small terrace remnant at the same altitude as the Traveller Creek terrace shows 2.5 m of well sorted pebble gravel which may be a more distal part of the same outwash surface.

Interpretation.

The gravel is proximal outwash deposited during the Traveller Advance. It is interesting to note that there is an absence of fine grained tills in all exposures of glacial sediments from the West Coast Range. This is primarily due to the dominance of resistant siliceous lithologies which do not produce much silt or clay when crushed. As a consequence the clasts and matricies of the sediments weather very slowly rendering relative dating methods based on weathering indices of little use.

6.6 The Fish Formation.

Introduction.

The Fish Formation like the Traveller Formation is a suite of glacigenic deposits derived from Mt. Jukes. It is limited in extent to the area south of the Governor River in the King Valley. Except at Baxter Rivulet, the Fish Formation unconformably overlies, and is indistinguishable from the Traveller Formation. The ice-contact sediments of this formation form relatively rare isolated low hills within low-lying outwash surfaces. The type section of the Fish Formation occurs at the Baxter Rivulet section at G.R. 875300 that was described previously.

Fish Creek section 1.

The Fish Creek section 1 at G.R. 870274 is exposed in an excavation in a low, hummocky terrace that consists of deformed ice-contact sediments and clast-supported diamictons (Fig. 6.32).

Description.

The diamicton is at least 3 m thick, is stratified in places and is massive elsewhere. Although the diamicton is poorly sorted, it is quite unlike other diamictons in the valley because it is clast-supported and generally lacks silt and clay. The sediment contains numerous small lenses and stringers of laminated and massive sandy silt. Some of the sand and silt lenses appear to be disturbed by a series of high-angle reverse faults. The positions of these faults are inferred from offsets in the laminated sediments (Fig. 6.32). The deformation appears to be syndepositional because the faults do not disturb all the sediments and appear to increase in intensity toward the top of the section.

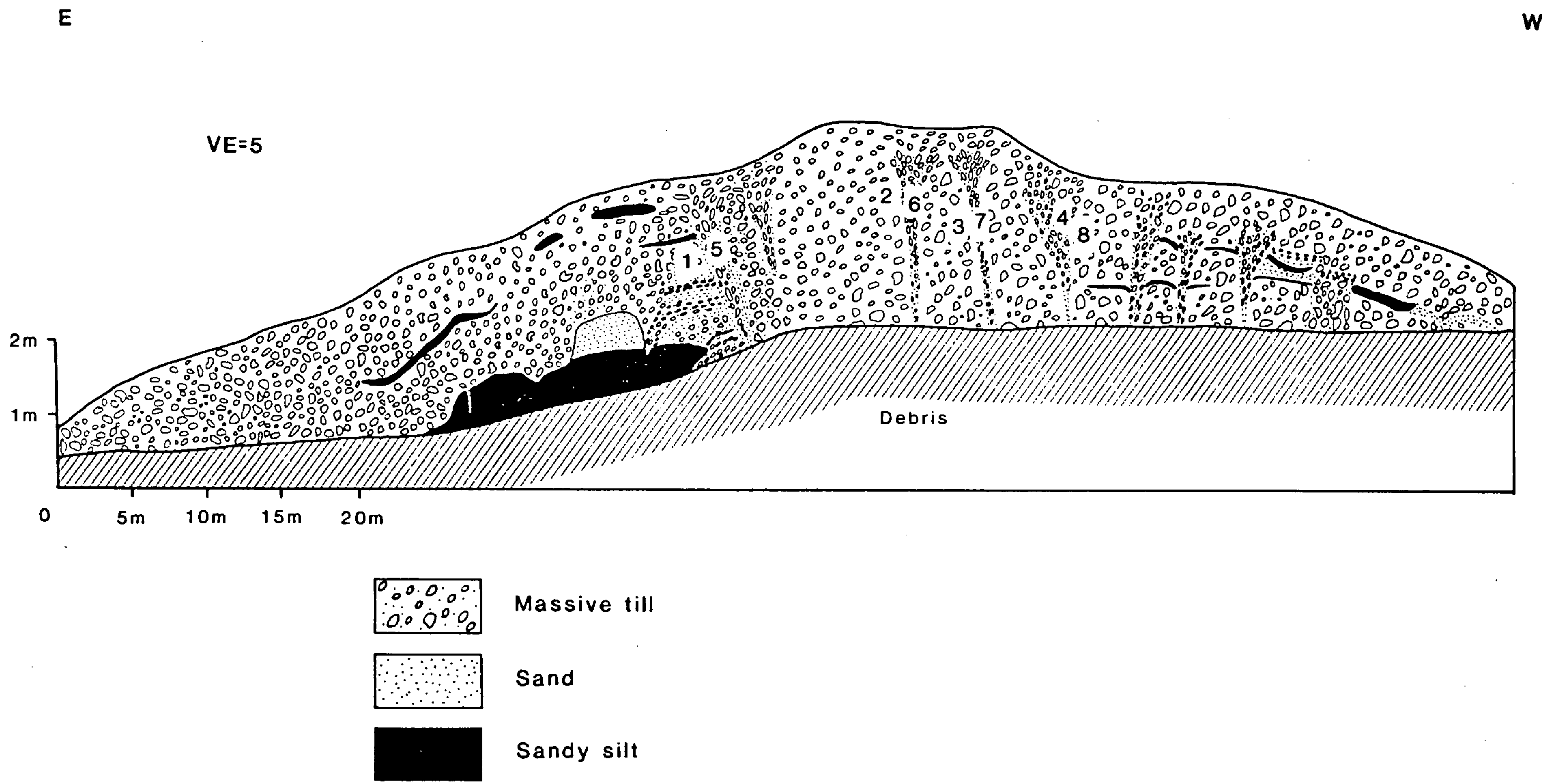


Fig. 6.32. Supraglacial sediments of the Fish Formation.

The lithology of the sediments consists of a mixture of locally derived Siluro-Devonian clasts, and Ordovician conglomerate and Cambrian volcanics derived from the West Coast Range (Fig. 6.32).

The pebble fabric of the diamicton is weak, and generally unimodal (Fig. 6.33). Fabric 1 has a bimodal orientation and fabric 4 has a girdle pattern that are not consistent with direct release from glacier ice. The strength of the clustering about the principle eigenvector is in the range of that typical for sediment flows (Fig. 6.34) as described by Lindsay (1968), and Lawson (1979a).

The section has several wedge-shaped clastic dykes and collapse structures that penetrate the diamicton by up to 3 m. (Figs. 6.35 and 6.37). These dykes have been described in a report that is reproduced in Appendix 2 where this group is referred to as swarm 2. The dykes are narrow, straight-sided and wedge-shaped structures, and many have walls made of vertically-dipping coarse sand (Fig. 6.35).

Unlike the clastic dykes in Thureau Formation sediments, the fills of these dykes cannot be distinguished from the host sediment by lithology or particle size (Figs. 6.32 and 6.36). The fills appear to consist of reworked host material that has slumped or fallen into open cracks. The pebble fabric of the host sediment is very different from that of the dyke fills (Figs. 6.33 and 6.34). The fabric of the fills show that clasts have a vertical dip that probably developed as the crack opened and filled. There appears to be a transition from wide and shallow collapse structures on the western side of the section to the narrow and deeper dykes in the central part of the section (Figs. 6.37 and 6.35).

Interpretation.

This section records deposition in a supraglacial ice-contact environment with syn- and post-depositional deformation as buried ice melted. Although the precise origin of the diamicton is uncertain, the intraformational lenses of silt and stratified sediment resemble subfacies 1, facies A

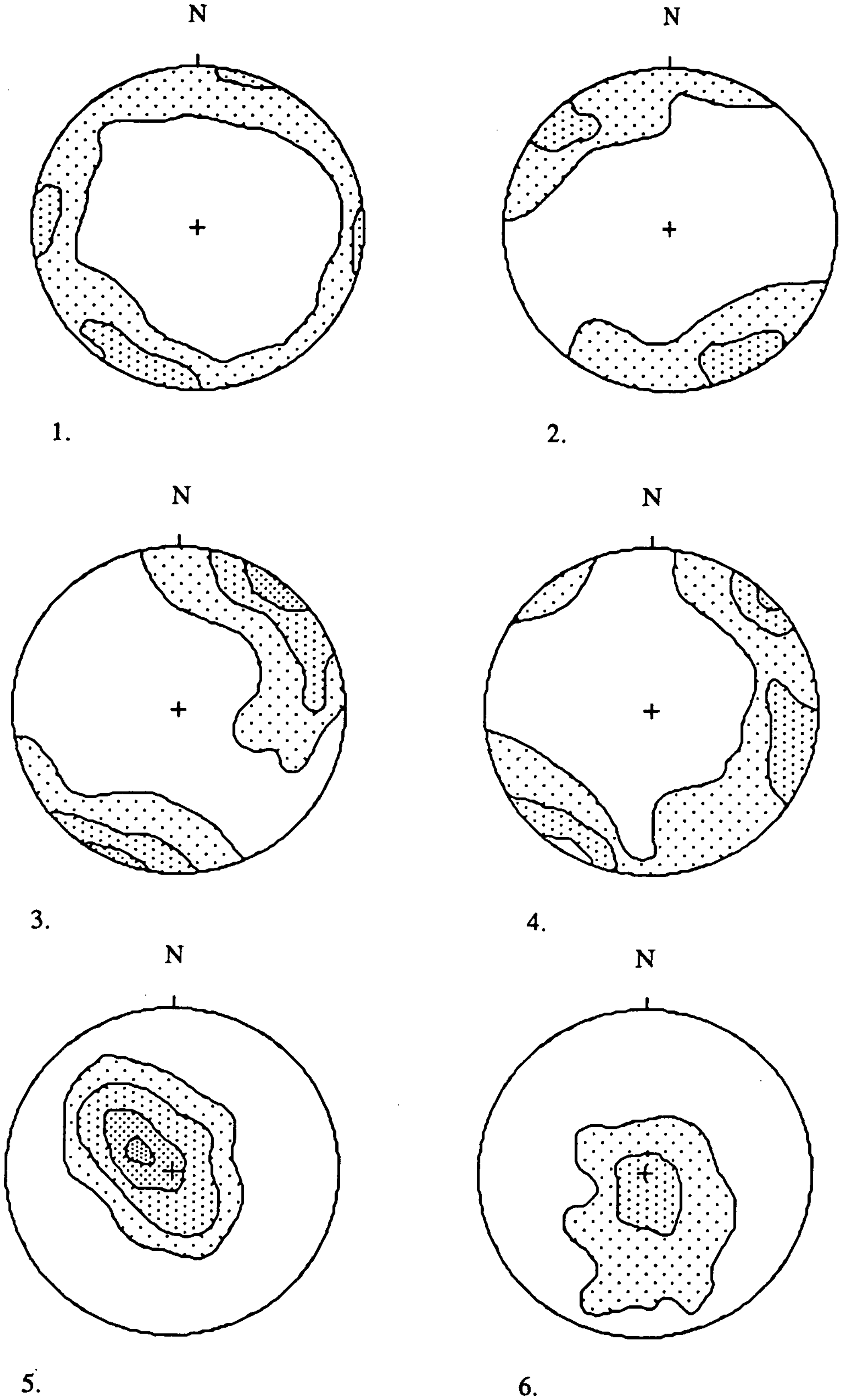


Fig. 6.33. Pebble fabric of the Fish Formation supraglacial sediments. See Fig. 6.32 for sample locations.

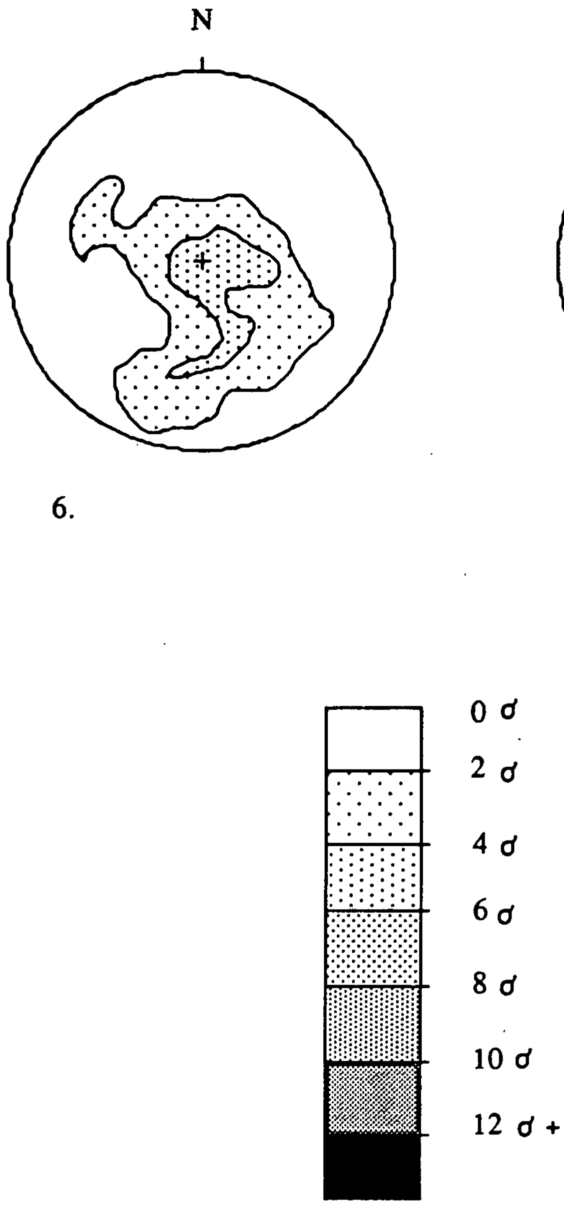


Fig. 6.33 (ctd). Pebble fabric of the Fish Formation supraglacial sediments. See Fig. 6.32 for sample locations.

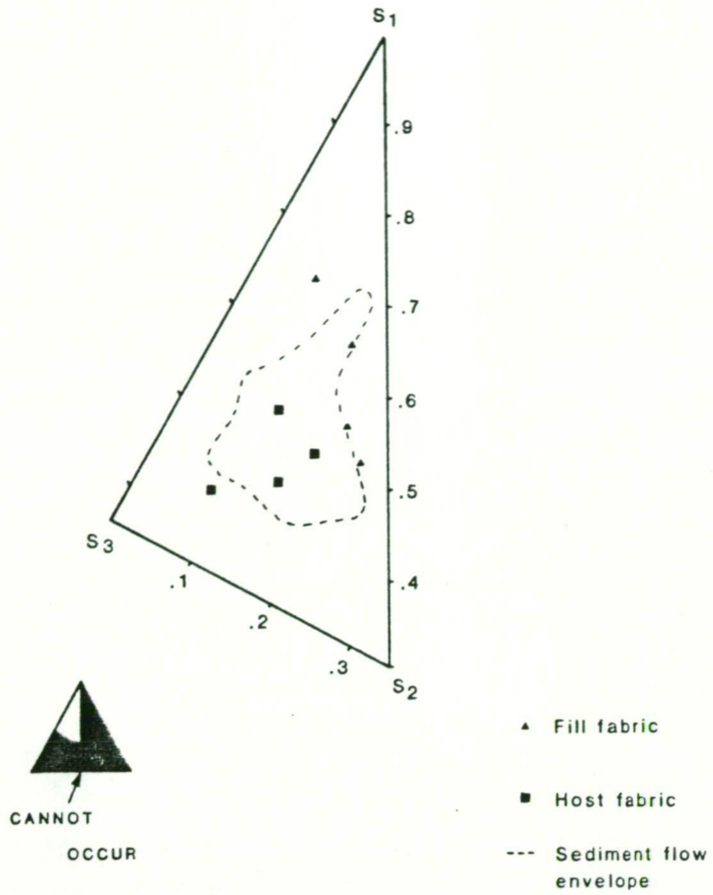


Fig. 6.34. Fabric strength of the Fish Formation sediments compared to those of sediment flows described by Lawson (1979a).

Fig. 6.35. Large straight sided clastic dyke in Fish Formation sediments.

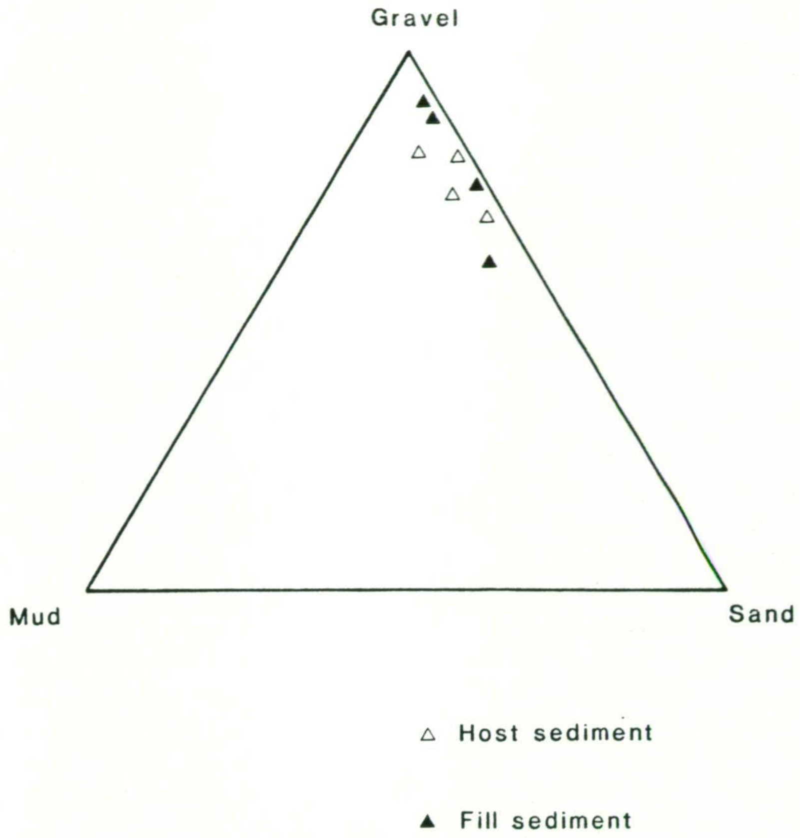


Fig. 6.36. Ternary diagram of particle sizes of host and fill sediments of clastic dykes in Fish Formation sediments.

Fig. 6.37. Collapse structures in Fish Formation sediments.

of supraglacial morainic till as described by Boulton and Eyles (1979). Such an association of deformed sediments is typical of deposition at the frontal margin of valley glaciers during retreat.

Silt and mud lenses, and stringers are common in supraglacial tills. They form as mud is eluviated from debris overlying ice and deposited at the ice-sediment interface. The position and geometry of silt stringers in the section suggest there are 2 or 3 ice sediment interfaces recorded in this section (Fig. 6.32). During the final melt of the underlying ice the silt stringers become increasingly deformed during the process that Eyles (1979) called backwasting. Both the deformation and multiple silt stringers are characteristic of supraglacial sediments.

The origin of the clastic dykes is not certain but they seem to be very similar to boulder-filled tension cracks that have been observed to form during the process of backwasting on Icelandic glaciers (ibid p. 1345). This process appears to be very similar to deformation in glaciofluvial gravels described by McDonald and Shilts (1975) using Sandford's (1959) model deformation experiments (Fig. 6.12).

The lack of silt and clay fractions in the till is partly because it is derived from the siliceous rocks of the West Coast Range and partly because it has been transported in a supraglacial position and has not undergone a phase of traction (Eyles and Boulton 1979).

Fish Creek section 2.

Two hundred metres west of the glacial deposits of the section described above, a terrace of Fish Formation outwash gravel is exposed in a road cutting at G.R. 868274. The section shows a thin diamicton lens resting on Devonian quartzite. It is overlain by sorted gravels and sands that were deposited in a braided river environment (Fig. 6.38).

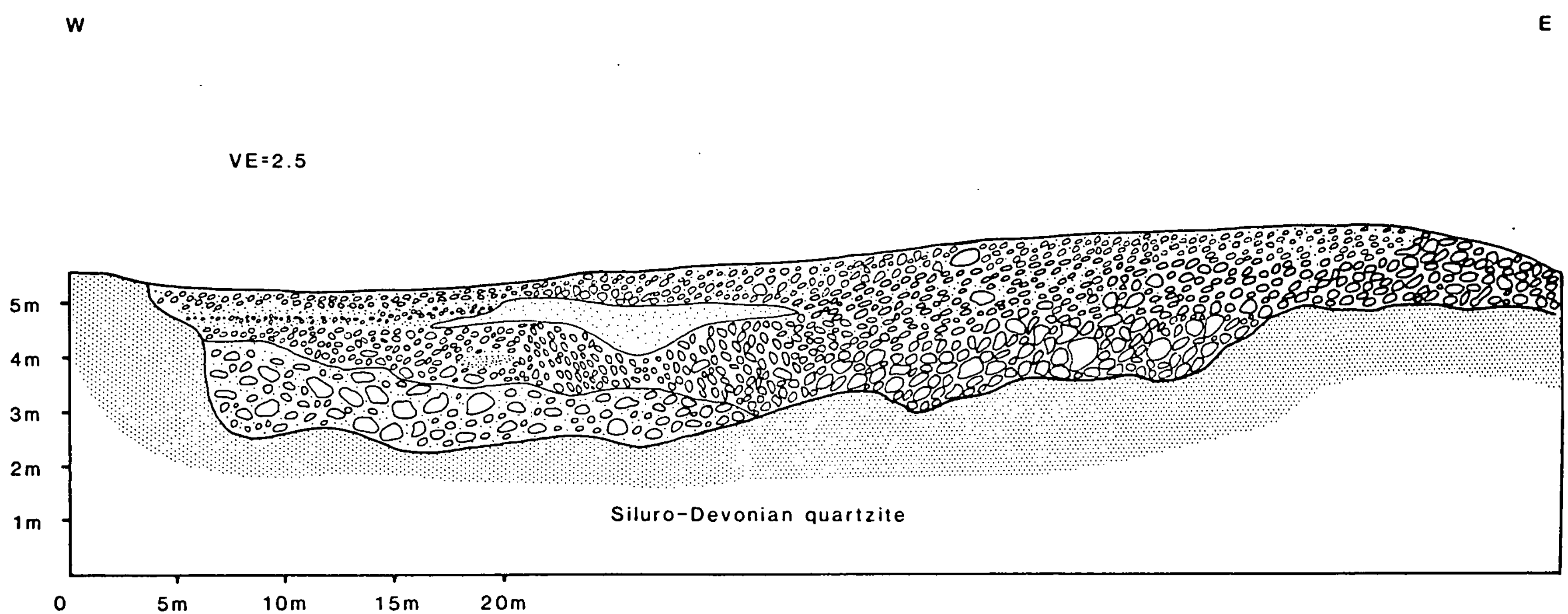


Fig. 6.38. Outwash sediments of the Fish Formation overlying an eroded till remnant (lower left) and resting on smoothed Siluro-Devonian quartzite.

Description.

The diamicton is similar to that shown in Fig. 6.32. It is poorly sorted, massive, partly iron-cemented and has a maximum particle size of 400 mm. The contact with the overlying alluvial sediments is eroded with the largest boulders from the till protruding through the contact into the outwash gravel.

The texture of the outwash gravel is highly variable. On the western side of the section the gravel is very coarse and appears to be a lag washed from the till. Above the lag the gravel has a strong imbrication and poorly developed horizontal bedding.

On the eastern side of the section there is a pronounced and uniform dip of c25° in the pebble fabric. The gravel appears to have been deposited on a bar face that migrated from east to west. The bar is 1 m thick, and is overlain by bedded fine pebble gravels and a lens of silty sand that may have formed when this channel was abandoned. There is a broad upward fining in the gravel.

Between this section and Figure 6.32 there are several shallow exposures of imbricated, coarse gravel interbedded with horizontally bedded sands and silts. One of these sections contains two clastic dykes in Devonian quartzite (Fig. 6.39). The fills of these dykes consist of rounded pebbles and cobbles from the overlying gravels and rare angular clasts of host rock. The quartzite dips nearly vertically and is brecciated, possibly by overriding ice. The dykes appear to occupy opened bedding planes that may have formed by preferential weathering and erosion, or by a glaciotectonic process. They are discussed in greater detail in Appendix 2.

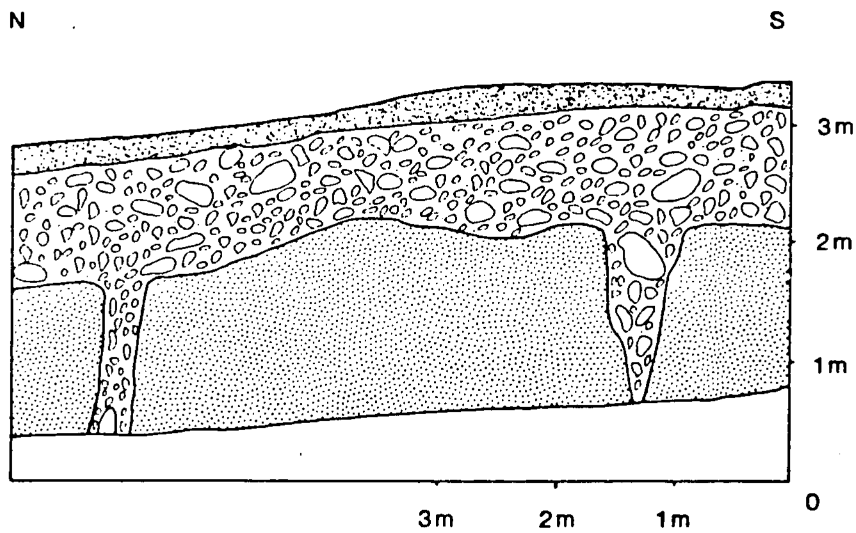


Fig. 6.39. Gravel dykes in Siluro-Devonian quartzite.

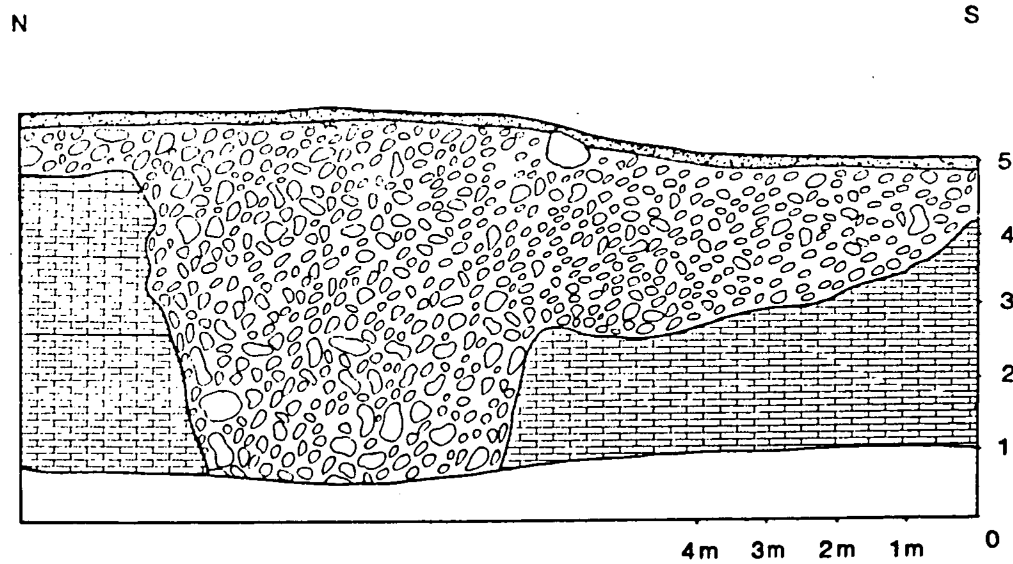


Fig. 6.40. Gravel dyke in weathered limestone.

Interpretation.

The sequence of gravels records deposition by a braided river over a partly eroded diamicton and Devonian quartzite. The direction of flow was from south to north, which in is toward the reader from Figure 6.38.

The section is typical of numerous shallow sections in Fish Formation outwash gravels. About 4 km south of these sections, near the Andrew Divide there are numerous sections of Fish Formation gravels which form a thin covering above black clayey silt (Fig. 6.41). The black clayey silt appears to be a local, transitional calcareous siltstone between Ordovician Gordon Limestone and Devonian Crotty Quartzite (F. J. Baynes pers. comm. 1985). Open tunnels and gravel lenses have been encountered in drillholes at depths of up to 14 m below the surface. In surface sections numerous steep-sided clastic dykes of coarse gravels penetrate the black clayey silts (Fig. 6.40). These dykes which appear to be filled solution tunnels and dolines are described in detail in Appendix 2.

The Andrew Divide appears to have had an extensive karstic drainage system that was filled by a flux of massive gravels that may have been from the Fish or an earlier ice advance. Since the filling, most of the karstic drainage system has not been redeveloped and the tunnels remain choked with coarse gravel in most parts of the valley. The only known surface karst features are at Dante Rivulet and in the adjacent Nelson River Valley (Kiernan 1980, 1983b).

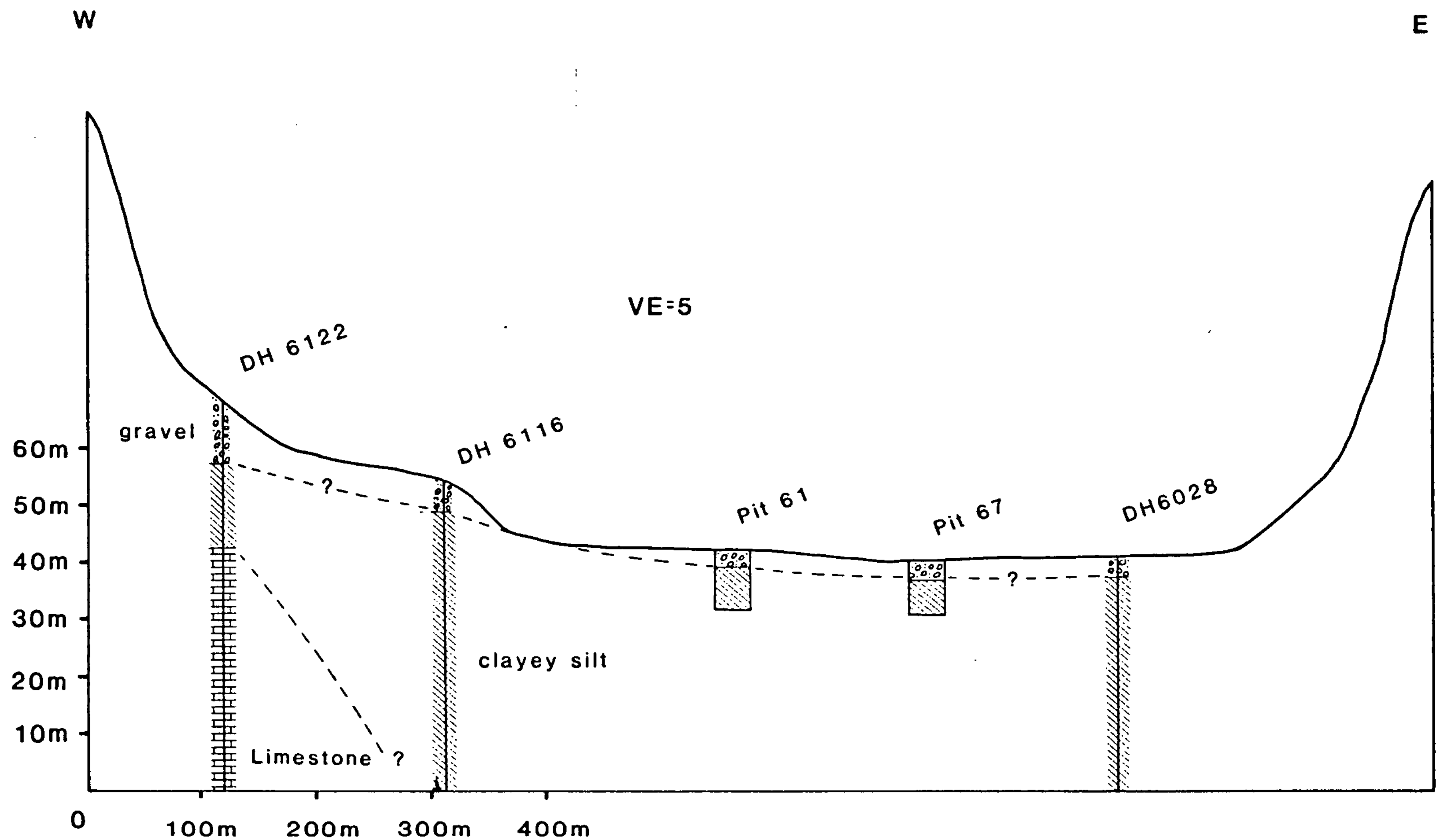


Fig. 6.41. Section across the Andrew Divide showing gravel resting on clayey silt (weathered limestone) and Ordovician limestone

6.7 The Governor Formation.

The Governor Formation sediments in the King Valley are limited in extent to a small area between the limits of the King advance and the southern limit of erratics (Fig. 2.1 and Map 1). The formation has no surface form because the sediments are eroded and buried by King Formation outwash gravel. Unconformities in these sections have been recognised from differences in weathering of the gravels. The differences suggest that the Governor Formation is at least twice as old as the King Formation.

The type section.

The Governor Formation type section is an excavation in a terrace 300 m east of Baxter Rivulet at G.R. 881303. The section consists of a 4.3 m upward-fining sequence of Governor Formation outwash gravels unconformably overlain by 3 m of bedded outwash gravels of the King Formation (Fig. 6.42).

Description.

Unit 1.

The Governor Formation consists of massive, coarse gravel with particle sizes up to 300 mm in diameter that is overlain by finer gravel interbedded with gravelly sand (Fig. 6.42). All the gravel is moderately well sorted, has numerous voids, is diffusely iron-stained to a bright orange colour and is locally iron-cemented. The lithology of the gravel consists of a mixture of clasts that have a West Coast Range, Eldon Range or local provenance (Fig. 6.43).

The differences in weathering rind thickness on Jurassic dolerite clasts suggests that a substantial period of time separates the deposition of the Governor and King Formations (Fig. 4.3 and section 8.2). The contact dips west at up to 10° and is marked by the protrusion of partly eroded

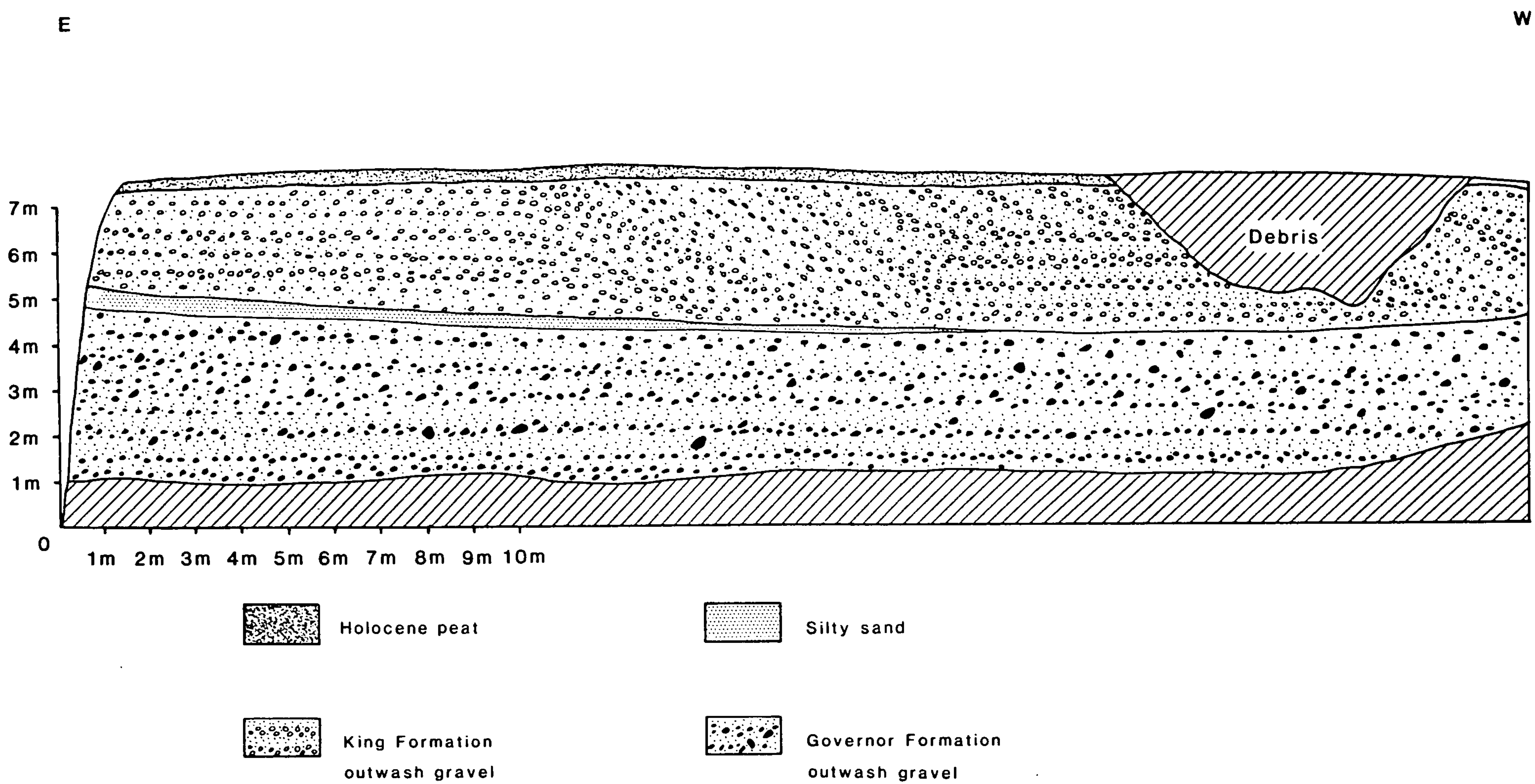


Fig. 6.42. Type section of the Governor Formation. Governor Formation outwash gravel is overlain by King Formation outwash gravel.

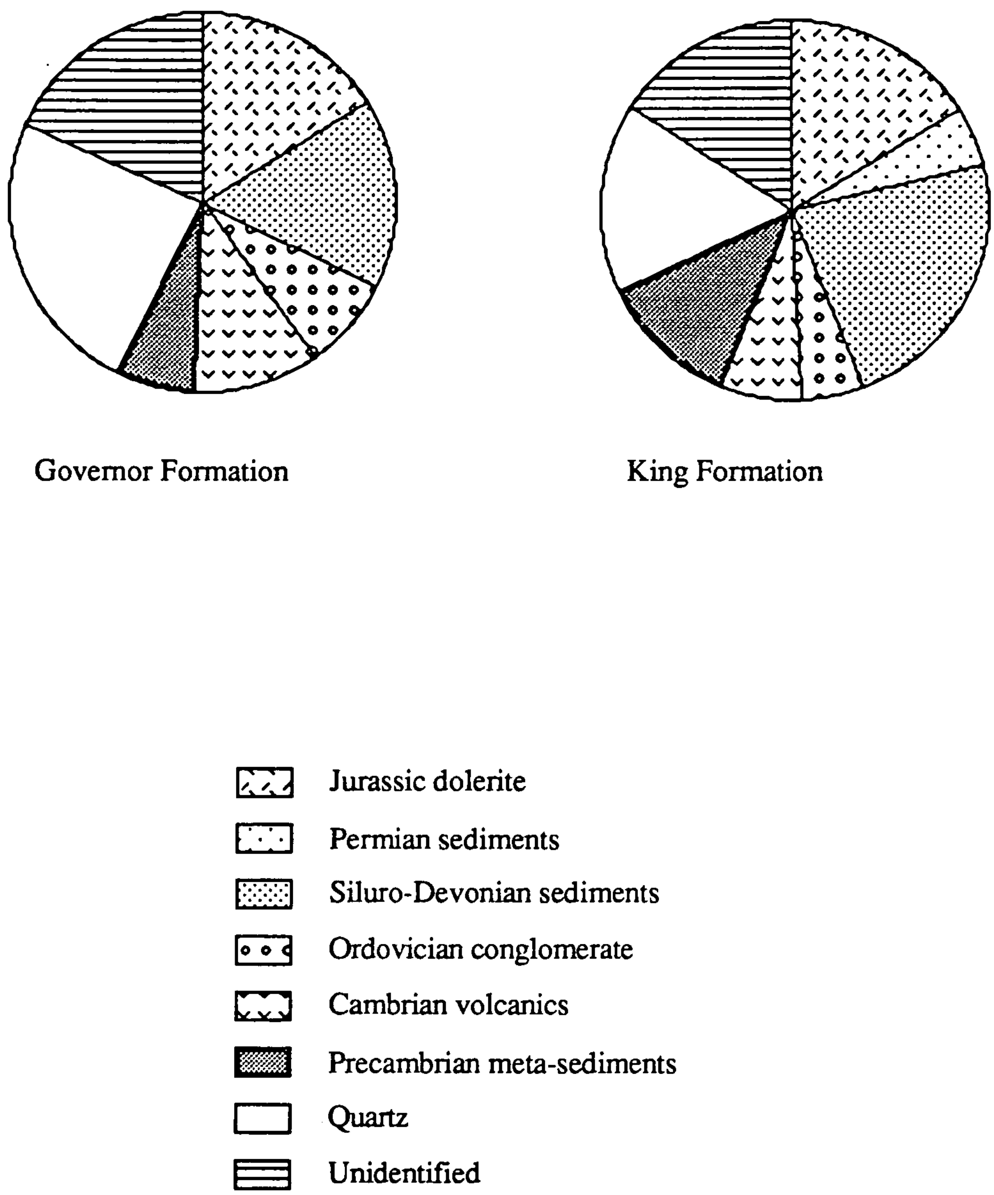


Fig. 6.43. Lithology of the Governor and King formations at the type section of the Governor Formation.

larger clasts through the contact and by dish-shaped load structures in a thin bed of fine silty sand. The deformation structures in the silt suggest that it was deposited during the onset of the King advance and was subsequently loaded by the gravel.

Unit 2.

The overlying King Formation gravel consists of coarse sandy gravel with openwork beds, pebbly coarse sands and coarse sands. The horizontal bedding on the eastern side of the section grades into trough cross-bedding in the central part of the section and horizontal bedding on the western side (Fig. 6.42). The lithology of the gravel is dominated by erratic Jurassic dolerite and Permian rocks (Fig. 6.43).

Both gravels are penetrated by steeply-dipping zones of humic staining from the overlying Holocene peat. The process of leaching of humic acids and deposition in vertical zones in sediments is common in the King Valley

Interpretation.

The section records sediments deposited in two different braided river environments separated by a substantial time period.

The Governor Formation gravels form an upward-fining sequence of braided river outwash gravels deposited in an ice proximal position. Lack of structure in the lowermost gravels suggests that they were deposited by a sheet flood. Because the texture of the gravel is coarser than the King Formation deposits they are thought to have been deposited close to the ice source. Although there is no surface form to this deposit the texture of the gravels at this section and at Baxter Rivulet (Fig. 6.27) suggest that the ice-terminal position was in a similar location to that of the King Formation (see Fig. 2.6).

In the King Formation gravels the transition from massive and crudely-stratified gravels to trough cross-bedded gravel is similar to the transition from bar-core to bar-face facies described by Eynon and Walker (1974). Trough cross-bedded gravel was probably deposited by avalanching of gravel down a bar face that migrated from east to west. The crudely stratified gravel from which the bar face developed is similar to bar core gravels which Eynon and Walker attributed to sheet flow during phases of aggradation. The slight upward coarsening of the gravels towards the terrace top is similar to the tendency Suggate (1965a) noted as characteristic of glacial outwash aggradation terraces.

The ages of both formations are discussed in Chapter 8.

6.8 The King Formation.

Outwash sediments of the King Formation form extensive terraces that extend from the middle to the lower King Valley, where the glacier terminated (Map 1). Because the ice advance terminated in a relatively narrow part of the valley, many of the ice-contact sediments have been eroded by outwash streams of the same and subsequent ice advances.

The type section.

The type section of the King Formation is a 28 m roadside exposure near the King bridge at G.R. 886314. The exposure is in a small terrace remnant high on the valley side that rests on Siluro-Devonian quartzite. Sediments in the exposure record an upward-fining sequence of proximal outwash gravels with large cut and fill structures that are overlain by locally derived slope deposits (Fig 6.44).

Description.

The lithology of the outwash gravels is a mixture of locally derived and erratic rocks (Fig. 6.44). The gravels have a much lower erratic content than other glacigenic sediments in this part of the valley. They have a variety of structures including horizontal bedding, cross-bedding, large cut and fill structures and old channel fills (Fig. 6.44). All the gravels are moderately well sorted, and are interbedded with sands and sandy gravels (Fig. 6.45). They rest on an uneven Devonian quartzite surface that is highly abraded and smoothed in some areas, and highly angular in others. Large angular blocks of up to 1 m diameter have been incorporated into the gravel in places (Fig. 6.44).

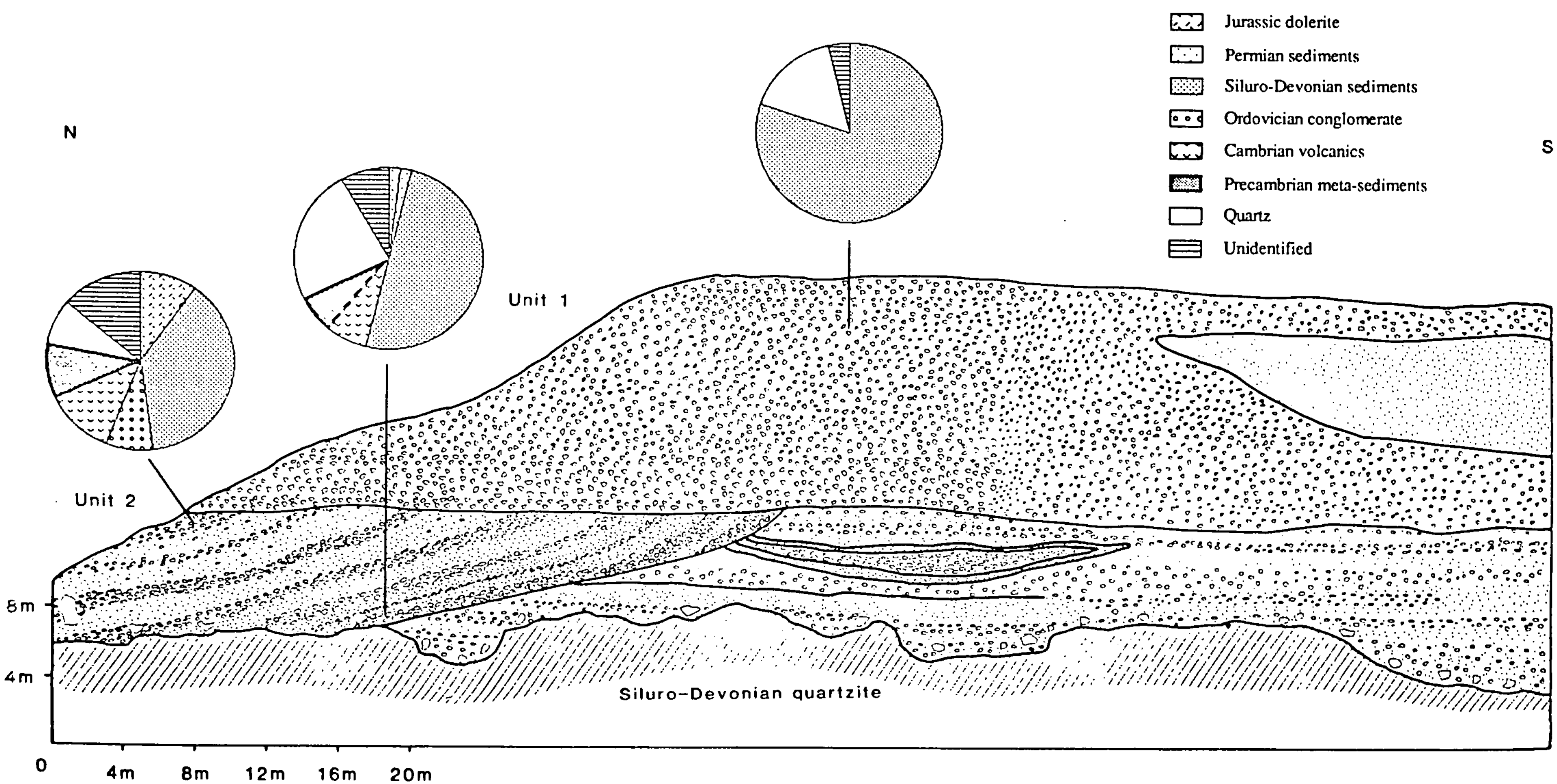


Fig. 6.44. Type section of the King Formation showing outwash gravel of a proximal outwash stream resting on Siluro-Devonian quartzite and overlain by locally derived slope deposits.

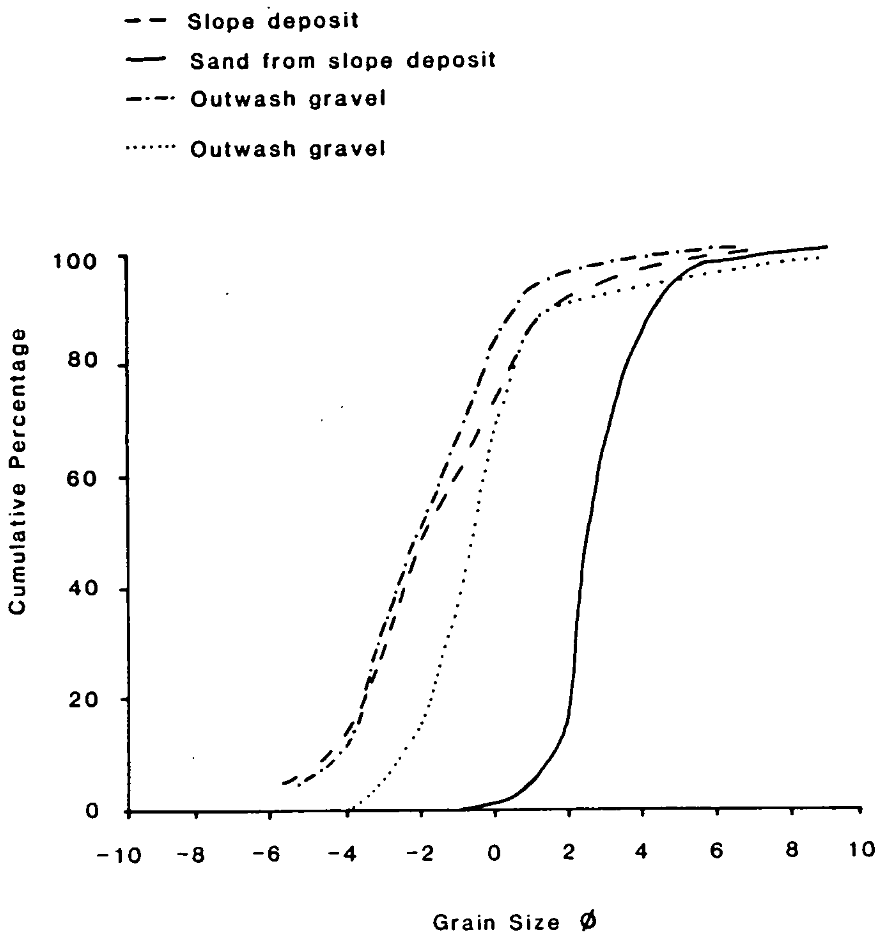


Fig. 6.45. Particle size distributions of sediments exposed at the type section of the King Formation.

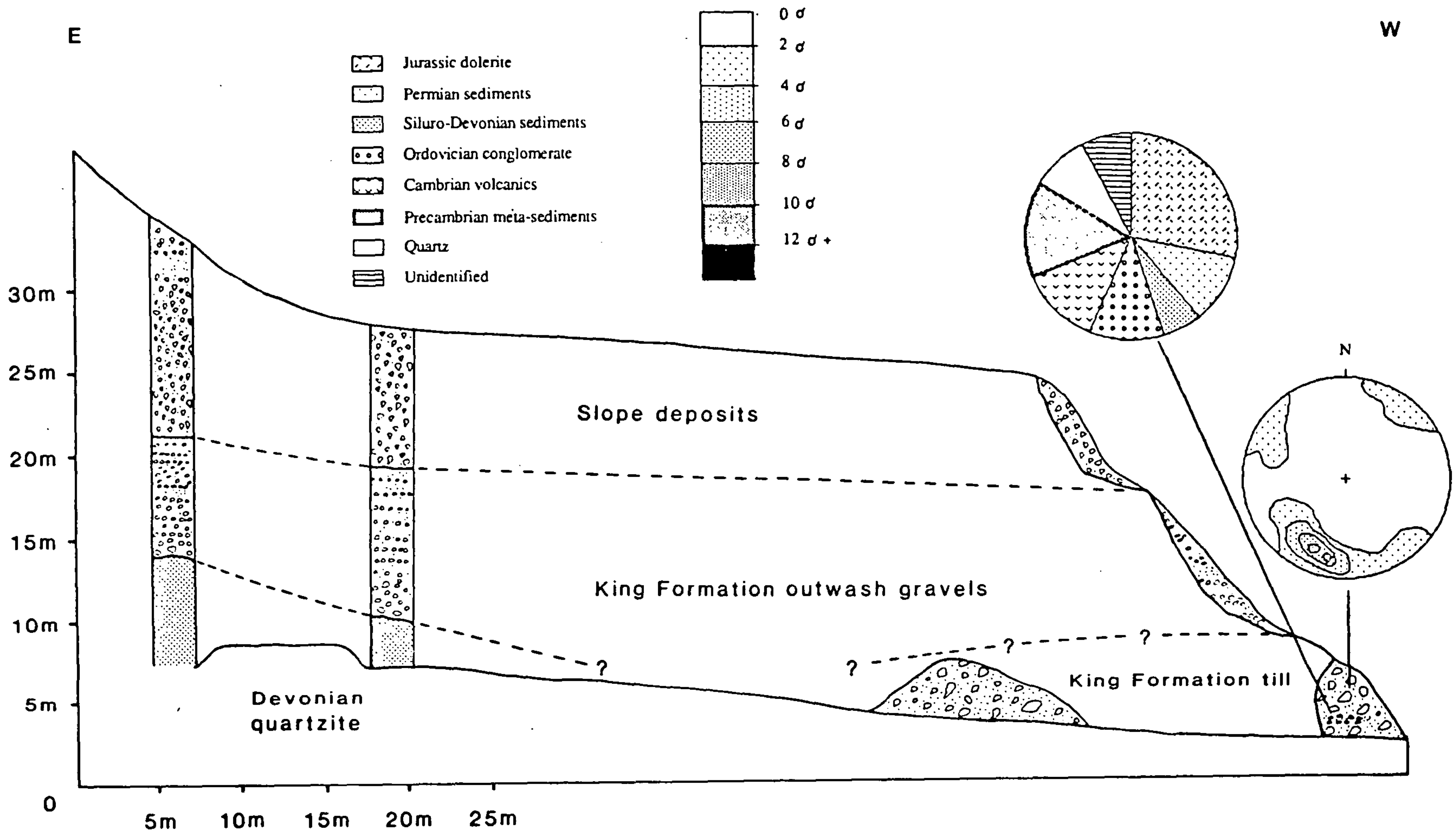


Fig. 6.46. Relationship between the King Formation outwash gravel and till that crops out 60 m west of the type section.

Up to 11 m of locally derived slope deposits overlie the outwash gravel and consist of angular clasts of Devonian quartzite with a sandy matrix. (Fig. 6.44). The quartz sand is well sorted and has the appearance of aeolian sands (Fig. 6.45).

Sixty metres west of the section a series of sections shows that a northward-thickening wedge of diamicton that lies between the outwash gravel and bedrock (Fig. 6.46). The diamicton is mainly massive but includes some semi-stratified zones that do not form discrete lenses. Both the weathering and lithology of the diamicton contrast with that of the outwash gravel. The weathering rinds on Jurassic dolerite in the till have a mean thickness of 7.6 mm and a standard deviation of 1.8 mm. The mean weathering rind thickness of clasts in the gravel is 4.7 mm and the standard deviation is 1.1 mm. The diamicton is considerably richer in erratic Jurassic dolerite and Permian sediments than in the outwash gravel (Figs. 6.44 and 6.46).

Interpretation.

The thick gravel deposit is proximal outwash that was deposited at the maximum of the King advance. The large cut and fill structures are from highly unstable stream beds that are typical of steep ice-proximal reaches of outwash fans. The diamicton buried in the outwash gravels is a melt-out till. The strong unimodal fabric, inclusions of stratified sediments and a massive structure are typical of melt-out tills which are generally difficult to identify (Haldrosen and Shaw 1982).

The geometry of the sediments suggests that the King ice advance terminated beyond the type section and that the wedge of till was probably deposited during retreat from the maximum position.

King River section.

Three hundred metres north of the King Formation type section at G.R. 878313 an exposure of a low mound shows ice-contact stratified sediments of the King Formation (Fig. 6.47). The glacigenic sediments are overlain by slope deposits and underlain by older tills, probably of either the Governor or Thureau Formation.

Unit 1.

The poorly exposed basal sediments are highly weathered and contain Jurassic dolerite clasts up to 40 mm in diameter that are completely weathered to clay. The pebble fabric of the diamicton shows a weak unimodal fabric dipping toward 122°. The gravel is probably an eroded remnant of the Governor or Thureau formations but it is impossible to say which.

Unit 2.

The weathered gravel is unconformably overlain by a series of deformed gravels, silts and diamictons with complex geometry. Some of the silts are intensely deformed (Fig. 6.47). Numerous small, high-angle reverse faults occur in the bedded sediments in the western side of the section. Sandy silts rest on a coarse diamicton on the eastern side of the section and on bedded gravels on the western side. The massive diamicton on the eastern side of the section is up to 1.5 m thick, contains lenses of poorly sorted, silty sand and has clasts up to 200 mm in diameter. It is clast-supported in places and matrix-supported in others. Its lithology is a mixture of erratic, West Coast Range and locally derived rocks (Fig. 6.48). The pebble fabric of the clast-supported part of the diamicton has a weak girdle pattern (Fig. 6.47).

The stratified gravel, which lies above and against the diamicton on the western side of the section, is poorly sorted, crudely bedded, imbricated, and contains clasts up to 100 mm in diameter. The pebble fabric of the dip direction of the A/B planes suggests it was deposited by a

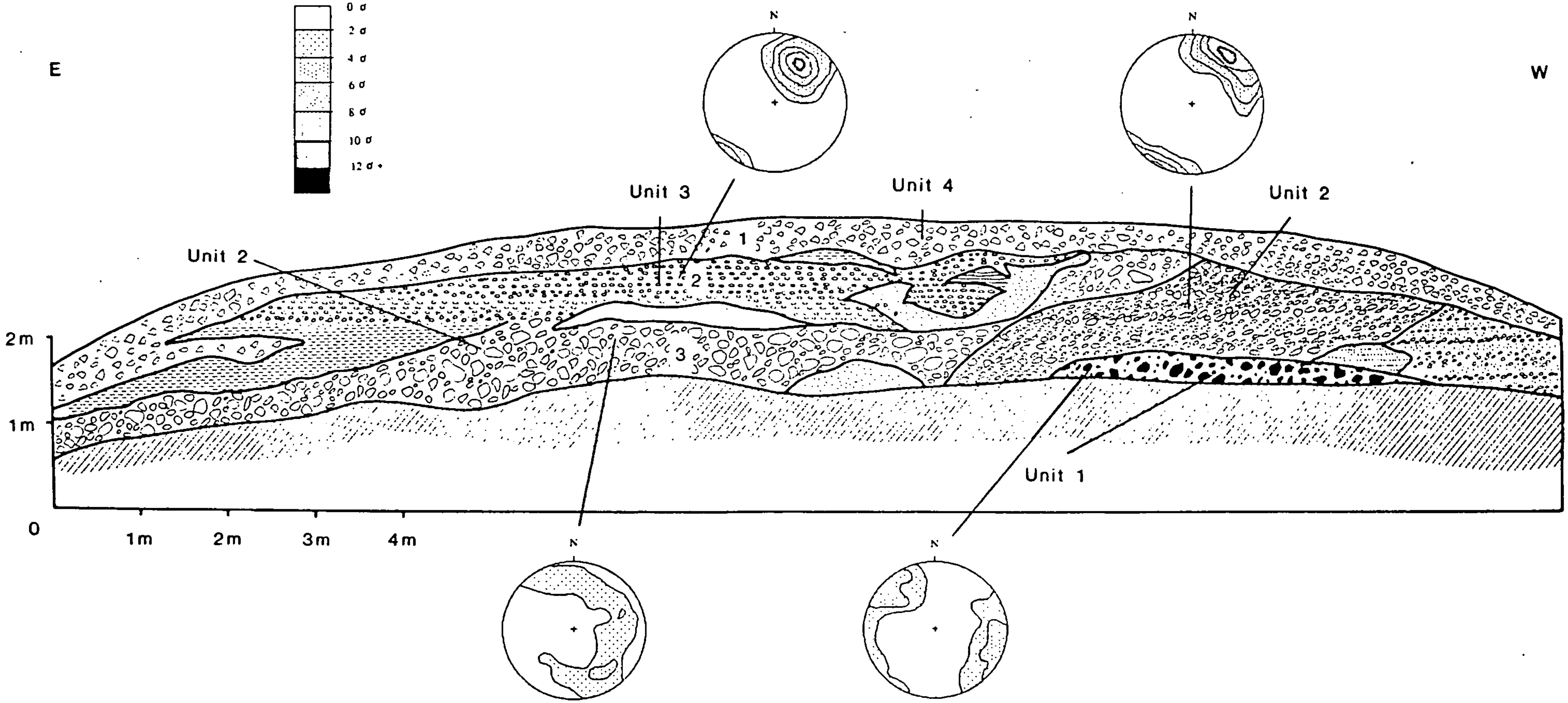


Fig. 6.47. Ice contact stratified sediments of the King Formation resting on Thureau Formation gravel (filled pattern) and overlain by slope deposits.

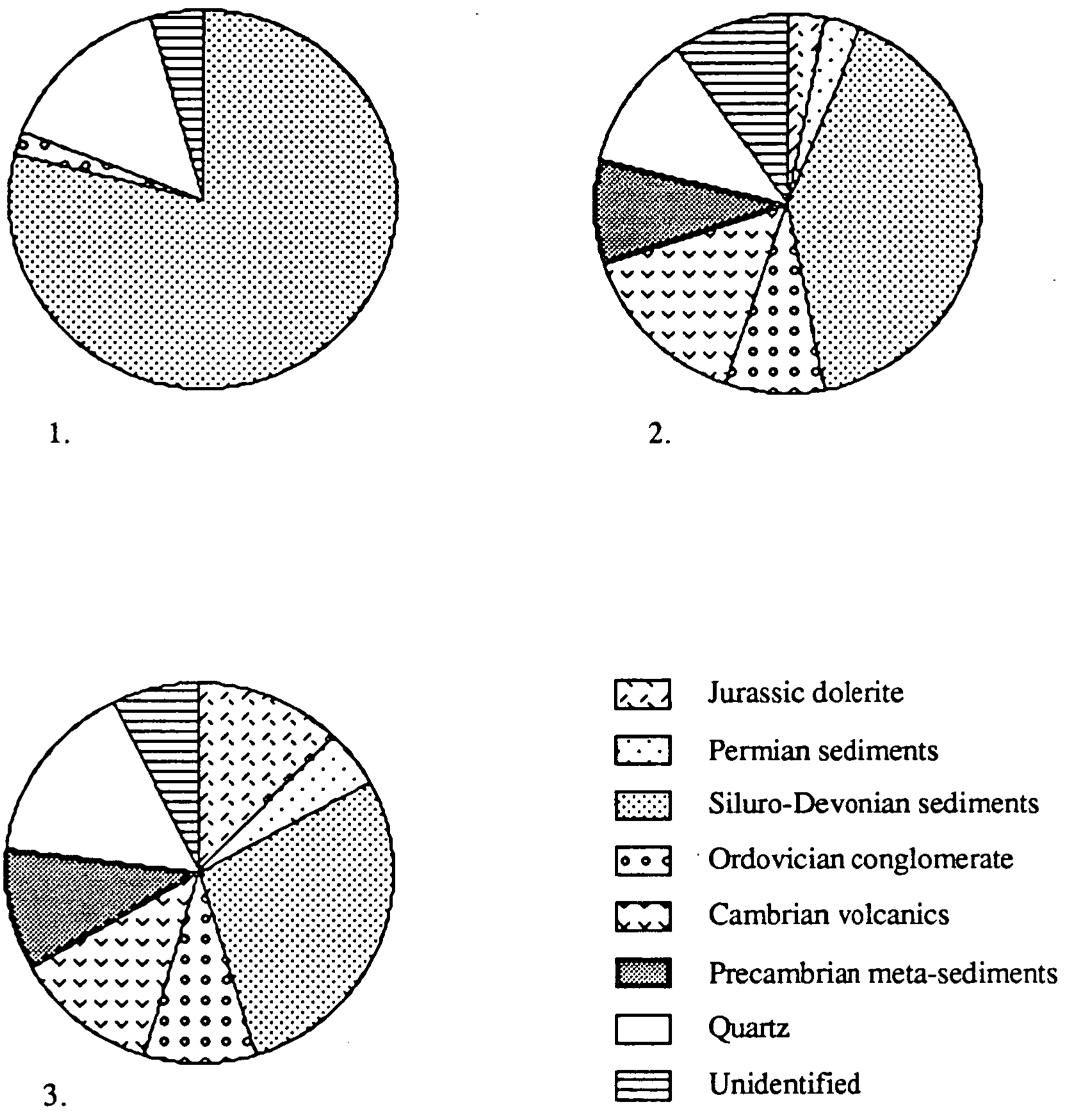


Fig. 6.48. Lithology of ice-contact sediments shown in Fig. 6.47.

current that flowed from the northeast to southwest. Parts of the gravel are highly deformed (Fig. 6.49) and appear to have small-scale collapse structures.

Unit 3.

Overlying the deformed diamicton, are well sorted, relatively undeformed, horizontally bedded gravel and silt deposits. The lithology of these gravels is less mixed than the underlying sediments and is dominated by locally derived rocks from the West Coast Range. There is also a relatively small contribution of erratic clasts from the Eldon Range. The pebble fabric of the dip direction of A/B planes of disc-shaped clasts shows a strong unimodal concentration very similar to that of the underlying imbricated gravel (Fig. 6.47).

Gentle warping of the bedding on the western side of the section suggests some degree of syn-depositional deformation. However, as the remainder of the gravel is undeformed it was probably deposited after the main period of deformation of the underlying sediments. The sharp contact with both the silts and diamicton suggests that a period of scour preceded deposition.

Unit 4.

The gravel overlying the imbricated gravel is entirely locally derived from outcropping Siluro-Devonian sediments (Fig. 6.48). The gravel is partly interbedded with the underlying glaciofluvial gravel and silt, suggesting they were deposited contemporaneously.

Interpretation.

This section records a supraglacial, deglacial environment in which syn-depositionally deformed glacial sediments are overlain by undeformed glaciofluvial sediments.

The girdle pattern of the pebble fabric of the diamicton is typical of that seen in the early stages of mudflows (Lindsay 1968, Lawson 1979a) and may have been deposited as a slumped or flowed

Fig. 6.49. Deformed silts resting on a coarse supraglacial diamictons in King Formation ice-contact sediments.

till. The interbedded and deformed nature of the relationship between the diamicton and outwash gravels is typical of supraglacial sediments.

The reconstructed sequence of events leading to the deposition of the sediment association seen at this section is:

1. melting of the glacier which terminated about 50 m from the section during the King ice advance;
2. deposition of sand lenses from subglacial tunnels onto the melting ice;
3. deposition of the massive diamicton on top of the sand, possibly by slumping from an ice surface;
4. deposition of gravels and sandy silt on top of the diamicton as melting proceeded;
5. melting of the underlying ice causing deformation during and after deposition;
6. deposition of the upper undeformed gravels and paraglacial slope deposits as the buried ice finally melts;
8. cessation of meltwater flow and burial of the glacigenic sediments by slope deposits.

The present surface form of the sediments as a small hummock is probably due to relief reversal of the deposits by accumulation in a basin on an ice surface similar to that described by Shaw (1972). However, it this instance the deformation that accompanies relief reversal is not evident because the it has of subsequent erosion of the margins of the hummock.

It is interesting to note that this section and several others like it throughout the King Valley do not have sediments of a subglacial origin which one would expect to be the basal sediments of each ice advance. In this section supraglacial sediments were deposited directly on sediments of an older formation, presumably subglacial sediments were eroded or not deposited.

Kelly Basin Road section.

Three kilometres north of the section described above at G.R. 885349 on the old Kelly Basin Road is a section that contains King Formation lodgement till overlying deformation till on the side of a large quartzite roche moutonnee (Fig. 6.50).

Description.

Unit 1.

The lower, unsorted diamicton consists of angular cobble-sized clasts of Devonian siltstone derived from the underlying rock and rounded quartz pebbles in a silty matrix (Fig. 6.50 and 6.51). In many cases these angular blocks have been dislodged from the siltstone and rotated small distances into the diamicton. The diamicton is moderately weathered and blocks of the less competent siltstone crumble when struck with a hammer.

The lithology of the diamicton is dominated by locally derived and West Coast Range rocks (Fig. 6.50) with an absence of erratic rock types. This suggests that there was very little mixing between these highly deformed sediments and the overlying diamicton that has a high erratic content.

The pebble fabric of the apparently randomly oriented blocks has a multi modal girdle pattern in one area and a weak unimodal concentration lower in the section (Fig. 6.53).

Unit 2.

The arc-shaped contact between the lower and upper diamictons is gradual and difficult to identify visually but within 10 cm there is a complete change in lithology from the upper diamicton containing erratics lithologies to the lower diamicton.

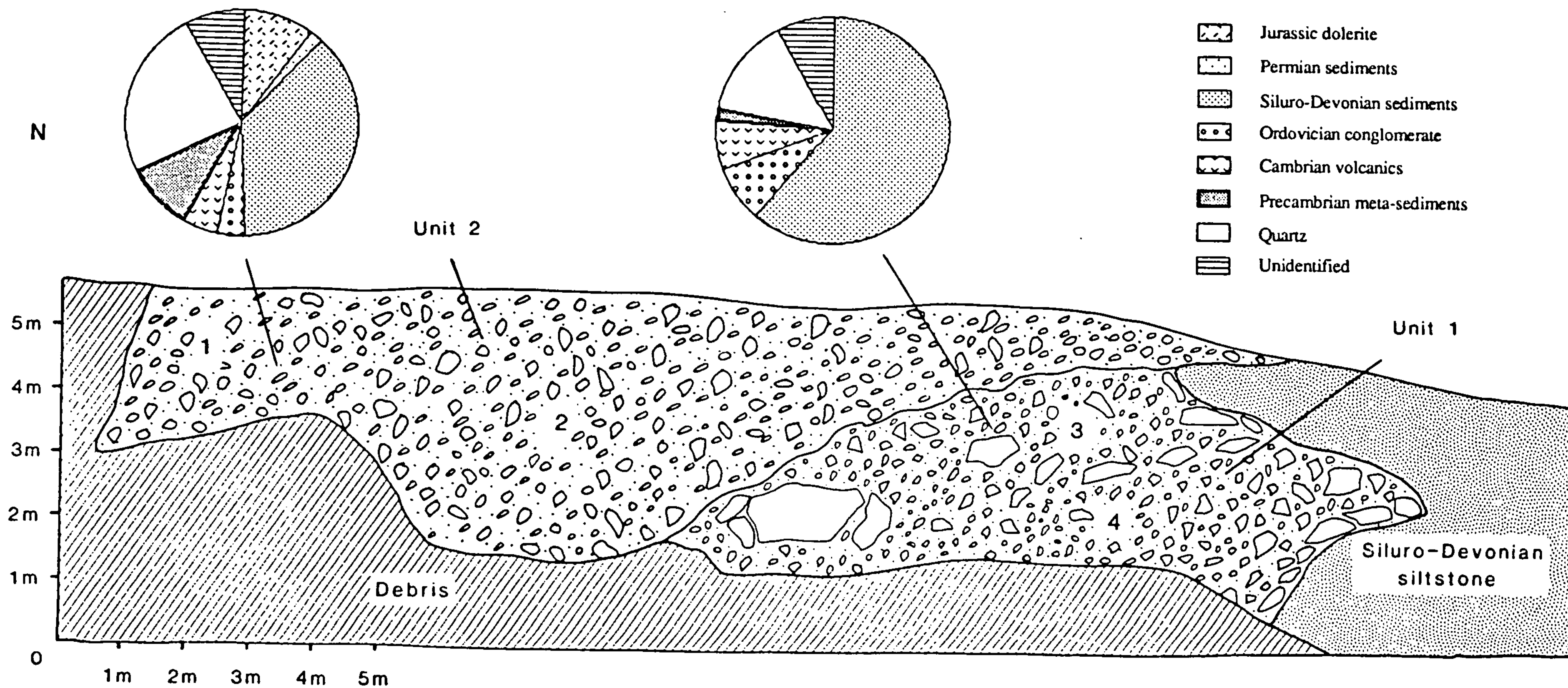


Fig. 6.50. Deformation till overlain by sheared lodgement till.

Fig. 6.51. Angular fragments of Siluro-Devonian siltstone and rounded quartz pebbles in a silty deformation till.

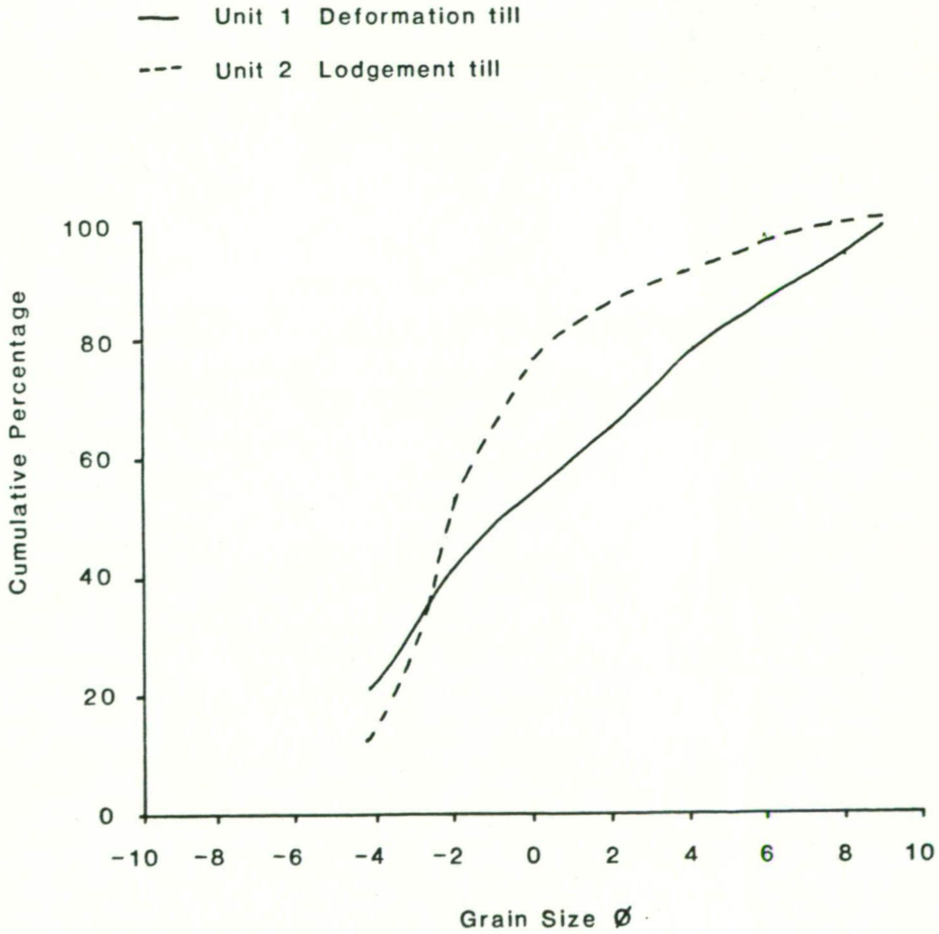


Fig. 6.52. Particle size distributions of the deformation till and subglacial lodgement till shown in Fig. 6.51.

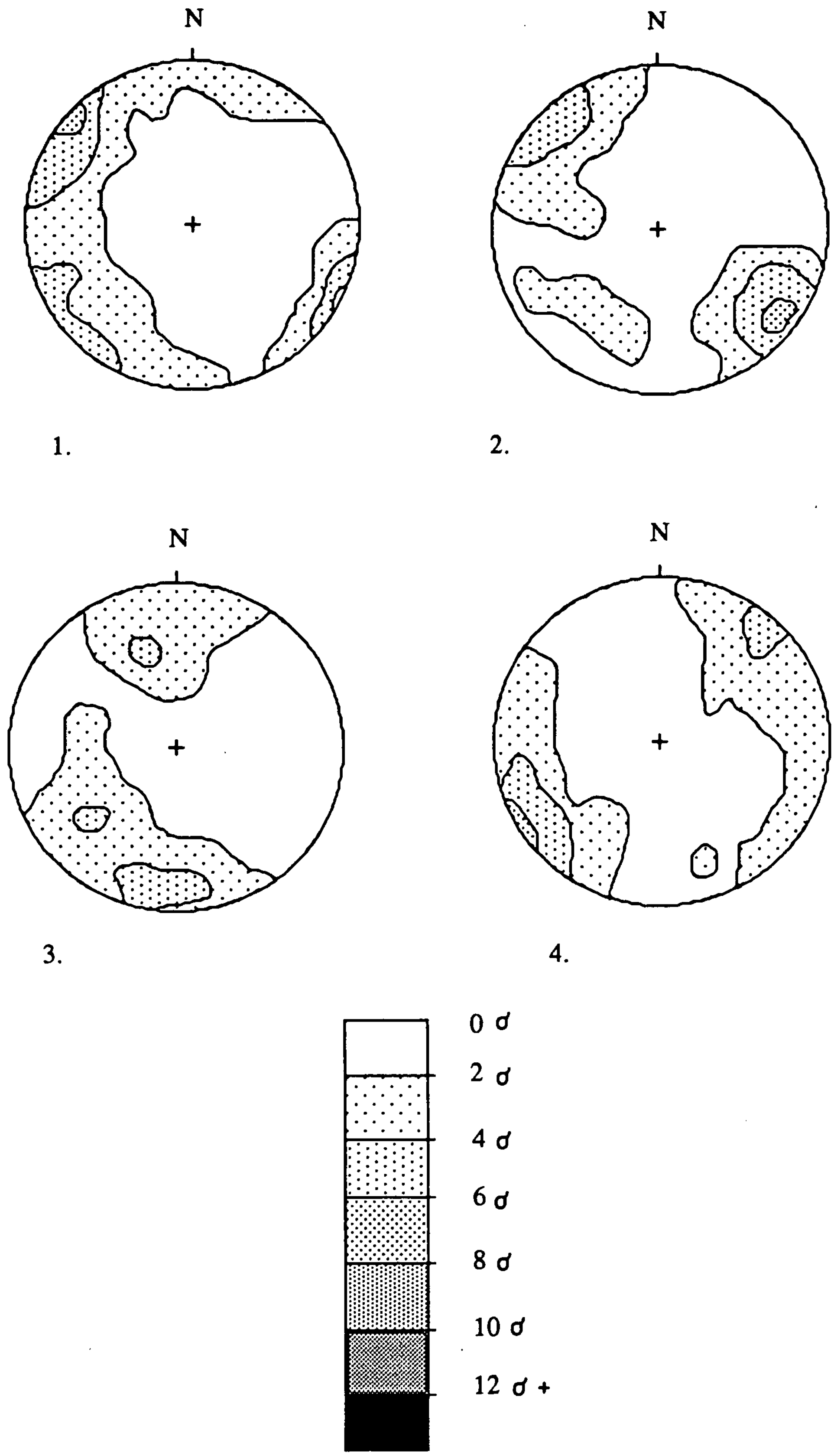


Fig. 6.53. Contoured equal area projections of pebble fabric of lodgement and deformation till. See Fig. 6.50 for sample locations.

The upper diamicton is up to 4 m thick, has a weak fissile structure and several weakly developed arcuate shear planes that are approximately parallel to the contact with the underlying diamicton. It is unsorted (Fig. 6.52) and has little textural variability throughout. The matrix is highly weathered and is iron stained.

The weathering rinds on Jurassic dolerite erratics have a maximum thickness of 35 mm, a mean of 14.8 mm and a standard deviation of 9.3 mm. The high standard deviation is unusual in weathering rinds of this thickness. A closer look at the distribution of rind thicknesses in the sample show that the sample has a bimodal distribution (Fig. 5.4). This is probably because two ages of sediments are represented here and the weathering rinds of the older deposit have not been completely removed during the reworking by the King Formation advance (see Chapter 5).

The lithology of the diamicton is dominated by the local rock type, Siluro-Devonian siltstone, but is a mixture of locally derived, West Coast Range and Eldon Range rocks (Fig. 6.50).

The pebble fabric is weak and bimodal, and has a stronger upglacier dip on the northern end (Fig. 6.54). The direction of maximum clustering (V_1) of the fabrics is within 10° of each other, and is parallel to the ice flow direction known from striae 300 m to the north of the section.

Interpretation.

The lower diamicton is a deformation till (Elson 1961), probably formed during an initial erosive phase of a glacial advance as relatively clean ice deformed a pre-existing fluvial sediment, plucked blocks of rock from bedrock and mixed them. The pebble fabric of the till probably reflects the stress fields during deformation. The overlying till was lodged from active ice as can be seen from the arcuate contact and shear planes. The weak pebble fabric at the northern end of the section is not entirely consistent with this explanation and suggests the till may have been deformed after deposition.

In a small section in a quarry 300 m north of the section described above, several wedge-shaped dykes penetrate Devonian quartzite and are filled with till. The dykes are narrow and irregular in outline and occur on the downstream end of a small roche moutonee. Three have inverted wedge shapes and rudimentary flow structures suggesting they were filled from above and from the sides. They were probably formed by the injection of basal till into fractures produced by the overriding ice. Appendix 2 contains a detailed description and discussion of the origin of these features.

Railway line section.

In the upper King Valley, between the limits of the King and Blackwood formations, extensive outwash terraces of the King Formation dominate the landscape (Map 1). The terraces are exposed in several locations and show coarse, poorly sorted, massive outwash gravels. The section described here (Fig. 6.54) is typical of these deposits and is exposed in an old railway cutting at G.R. 888347.

Description.

The lower 3 m of the section consists of interbedded sands, fine gravel and laminated silt (Fig. 6.54). All are diffusely iron stained and the gravels contain numerous iron pans up to 5 cm in thickness. The sediments consists of a sequence of laminated silts, massive and cross-bedded sands and gravels overlain by 2.7 m of coarse, poorly sorted gravel (Fig. 6.54). The contact between the coarse gravel which has clasts up to 500 mm in diameter, and the bedded sediments is sharp and appears to have been eroded. The coarse gravel is generally massive and has little fine sand or silt.

The pebble fabric measured as the dip direction of A-B planes shows a weak imbrication suggesting the gravel was deposited by a flow from southeast (Fig. 6.54). The lithology is a mixture of locally derived, West Coast Range and Eldon Range rocks.

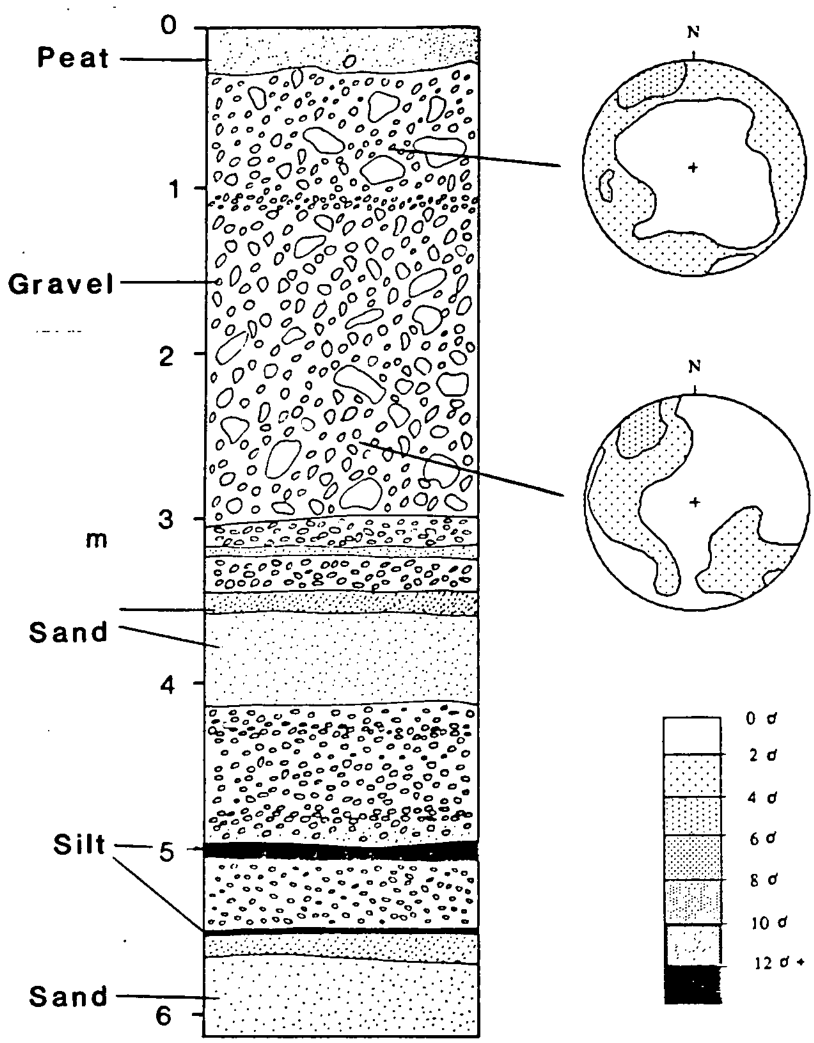


Fig. 6.54. King Formation outwash gravels in the middle King Valley, 3.5km north of the King Formation type section.

Interpretation.

The position of the terrace between two known glacial limits and above the aggradation level of the Blackwood Formation suggests that at least the top of the terrace was deposited during the retreat of the King Formation. It is not a aggradation terrace because only the upper metre of the surface was deposited by meltwater.

6.9 The Nelson Formation.

Introduction

The Nelson Formation forms a thick sequence of laminated silts that lie between the King and Blackwood formations. The type section is known from drill log data and excavations through the Blackwood Formation end moraine at G.R. 896370.

Unit 1.

The gravels at the base of the sequence are known only from drill cores and are not exposed. They consist of fine gravels with a maximum particle size of 100 mm and a mixed lithology that is dominated by locally derived Siluro-Devonian sediments. In some drill cores there appears to be a lag of large boulders at the base of the gravel just above the rockhead (J. Geidl pers. comm. 1986, Fig. 6.57).

Unit 2.

The contact between the gravels and the overlying laminated silts is known only from the drill cores. Although the character of the contact is difficult to determine it appears to be sharp. The silts consist of pale green beds with multiple sublaminae less than 1 mm thick, and occasional dark grey laminae up to 40 mm in thickness (Fig. 6.55). According to the classification of Folk *et al.* (1970) the dark green laminae are silts and the pale green laminae are muds (Fig. 6.56).

The silts are relatively uniform throughout the section to within 4 m of the contact with the overlying diamicton where there are intensely deformed. Just below the contact the silt is brecciated, and dark grey pieces of silt are chaotically dispersed in a massive green matrix. Within 1.5 m above the brecciation the intense deformation gives way to gentle warping of the laminae, and numerous small, high-angle reverse faults occur. Figure 6.55 shows some of the reverse

Fig. 6.55. A low angle reverse fault in laminated silts of the Nelson Formation.

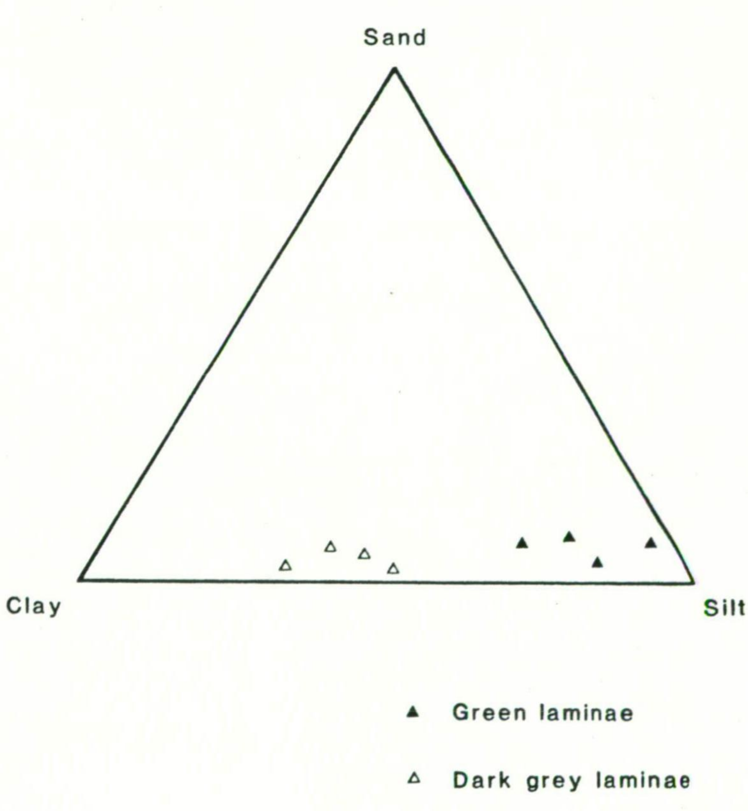


Fig. 6.56. Ternary diagram of the particle size of the deposits shown in Fig. 6.55.

faults which typically have throws of around 50 mm and up to 170 mm. A further 2 m below this the silts are not faulted but are gently warped and dip west at up to 17°.

The remainder of the silts have no other primary or secondary structures. Of particular note and unusual for this kind of deposit in the King Valley is the absence of dropstones. Several samples were examined for fossil pollen but only three grains of Compositae were found though the processed samples had a great deal of organic debris (G. van de Geer pers. comm. 1986). However, two unidentified seeds were found buried in the upper part of the silt sequence at 200 m altitude.

Numerous, flat, carbonate-cemented concretions occur in the dark grey silty laminae at around 190 and 210 m altitude. The origin of the carbonate is not known but similar concretions are common in glacial lake sediments in New Zealand.

A seismic reflection survey of these unconsolidated sediments encountered several shallow-dipping reflectors within the silts (Robertson 1986). Although Robertson interpreted the reflectors as possible evidence for sand lenses in the lake silts, none of the cores or excavations encountered such lenses. The reflectors therefore may be related to the presence of horizons of carbonate concretions associated with the dark laminae.

Interpretation.

The gravel that underlies the silts is probably a remnant of the King Formation.

The overlying laminated silts were probably deposited in a deep lake as turbidity currents settled out. The temporal relationships of the individual laminae are unclear but investigations of sedimentation in temperate proglacial lakes suggests it is unwise to call such laminae varves and interpret the sedimentary couplet as the product of one years sedimentation.

A lake of the size and depth necessary for the formation of sediments of this type must involve a large dam in the lower part of the King Valley. The most likely origin for a dam of this sort is damming by a moraine of the previous King advance. The geomorphology of the valley suggests that this is most likely to occur about 1.8 km upstream of the Governor River confluence at the southwestern edge of the Thureau Hills where there is a constriction in the valley that could easily be blocked (Fig. 2.1). The absence of pollen in the silts suggests that they either accumulated very rapidly or in a cold environment. The long, uninterrupted record of silt deposition, the absence of dropstones, scour, sediment flows and structures typical of ice-marginal lakes as described by Shaw (1975), suggests that the depositional environment was remote from direct glacial influences and without floating ice. The silts are therefore interpreted as representing an interstadial period separating the King and Blackwood advances. The length of the interstadial can be estimated assuming similar sedimentation rates to Holocene proglacial lakes in New Zealand. Pickrill and Irwin (1983) calculated the sedimentation rate in Lake Tekapo as 1.0 ± 0.1 cm.yr.$^{-1}$, using this rate the 43 m of silts known from the Blackwood type section could have accumulated over a period of $4{,}300 \pm 430$ yrs.

The interpretation is equivocal. An alternative explanation is that the silts accumulated when the climate was glacial and ice was absent from the valley but occupied the Tyndall Plateau and higher peaks of the West Coast Range.

Regardless of climatic inferences, the silts clearly represent a significant temporal break in direct glacigenic deposition and lie between two glacial advances. On these grounds alone the deposit can be called a stratigraphic interstadial.

6.10 The Blackwood Formation.

Introduction.

The Blackwood Formation forms extensive outwash terraces and a large arcuate moraine ridge in the middle part of the King Valley (Map 1). In the lower King Valley near the Governor River numerous low terraces are inset into older King Formation deposits and are interpreted as being reworking associated with outwash streams of the Blackwood advance. Other large exposures of glacigenic sediments occur at the mouth of Linda Creek and near the Nelson River at Valley Creek.

The type section.

The type section of the Blackwood Formation consists of two 3 m-deep trenches that run from a terminal moraine crest at 242 m to the King River at 194 m, and numerous drill holes that penetrate up to 80 m of Quaternary sediments resting on Siluro-Devonian siltstone (Fig. 6.57).

The deposits exposed in excavations form a sequence of sediment flows, outwash gravels, massive diamictons and lodgement till deposits. The sequence rests on the Nelson Formation and records the advance of the King Glacier beyond the end moraine, followed by formation of the moraine on a proglacial outwash surface. The near surface sediments have been affected by post-depositional mass movement. This has caused intense local plastic deformation, faulting and brecciation of the laminated silts, and formation of clastic dykes as fills of tension cracks in the overlying tills and outwash gravels (Appendix 2).

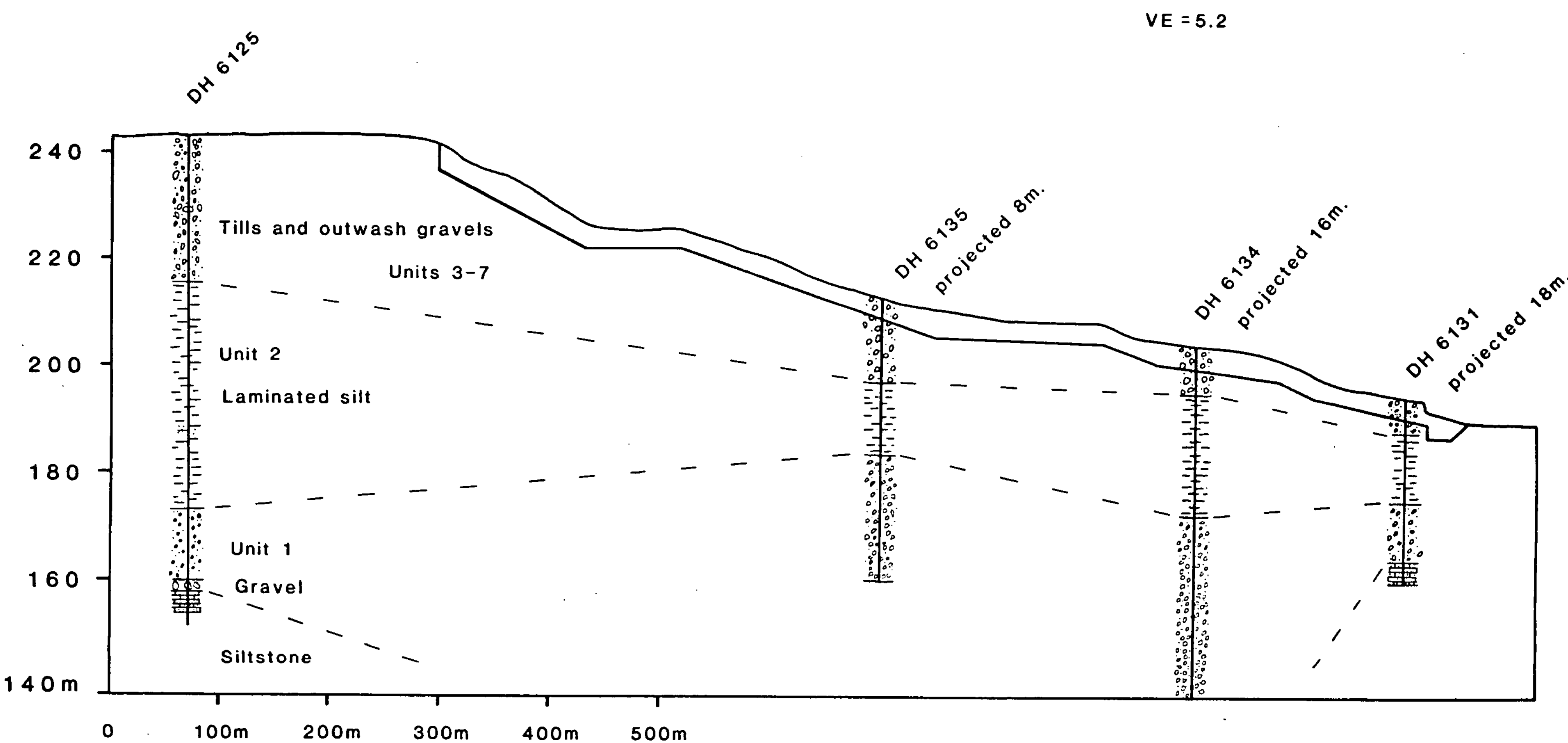


Fig. 6.57. Type section of the Blackwood Formation showing the relationship between Blackwood Formation sediments (Tills and outwash gravel) and the Nelson Formation (laminated silts) and the King Formation (gravel?) which rest on Siluro-Devonian siltstone.

Description.

In the following description, units 1 and 2 form part of the Nelson Formation and were discussed in section 6.9.

Unit 3.

The contact between the silts of the Nelson Formation and the overlying diamicton is sharp and eroded in most places, but at two locations in the trenches pale green well sorted medium sands form what may be a transitional and conformable deposit. There appears to be a rapid change in the energy level of the environment, probably associated with the onset of the Blackwood advance.

The diamicton is massive, unsorted, and has a maximum particle size of 300 mm. The lithology consists of a mixture of locally derived West Coast Range and Eldon Range rocks. At 12 m and 14 m thin beds of laminated silt and crudely bedded gravel dip west at 17° (Fig. 6.58). For the remainder of its thickness the deposit is massive but occasional small sand lenses occur.

Unit 4.

The overlying 2.2 m-thick diamicton has a pronounced fissility, is highly consolidated, and consists of pebbles up to 50 mm in diameter supported in an iron-stained silty matrix. The pebble fabric is strong, unimodal and is parallel to the reconstructed ice flow direction which was from north to south (Fig. 6.58). The lithology of the diamicton is similar to that in the overlying and underlying sediments.

Unit 5.

The diamicton overlying the fissile diamicton is very similar to unit 3. It is massive, has a maximum particle size of 300 mm, a mixed lithology and a weak unimodal pebble fabric (Fig. 6.58). Where the diamicton is exposed at the surface in the trenches, several narrow wedge-

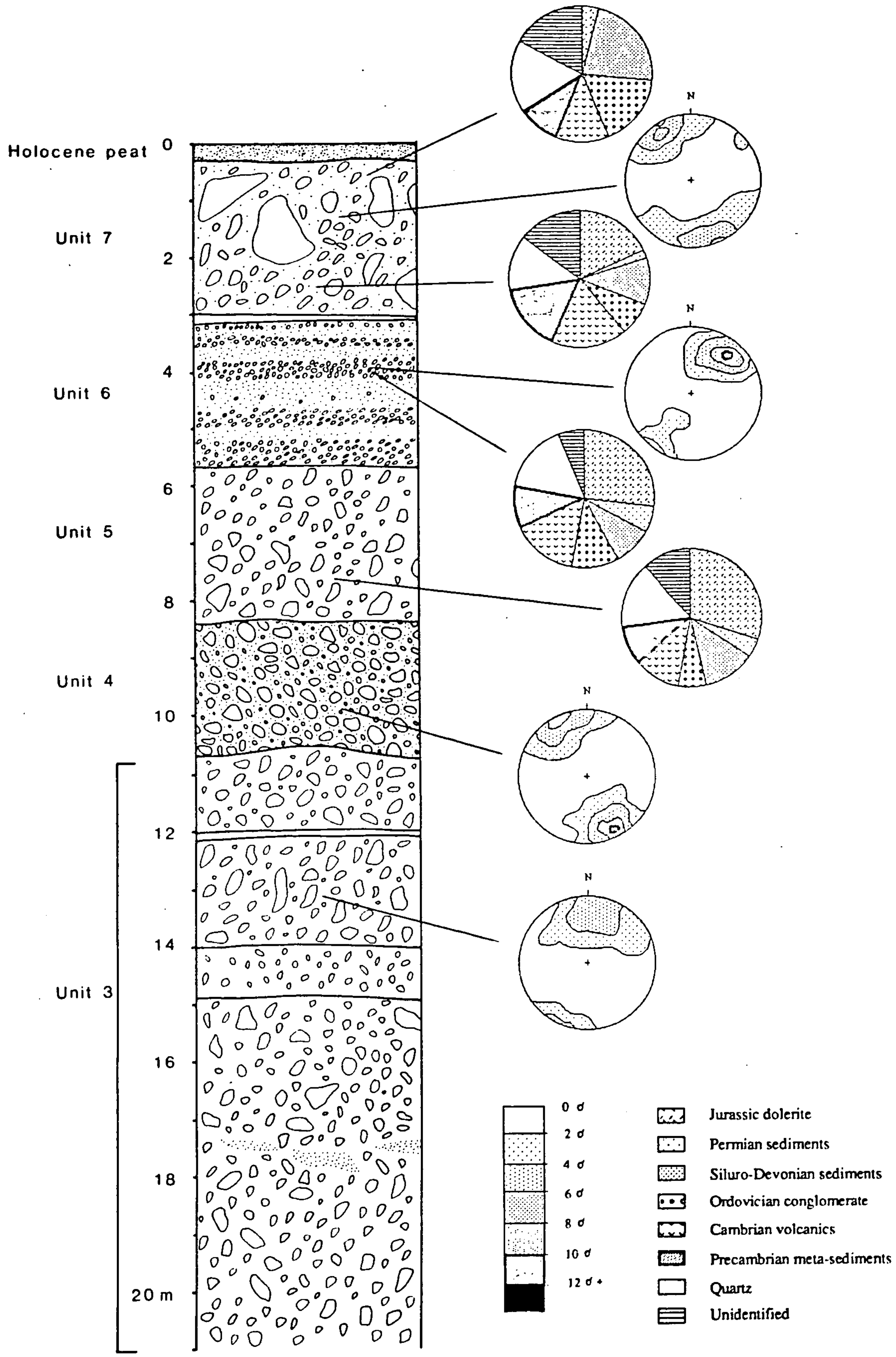


Fig. 6.58. The sequence of sediments exposed in the Blackwood Formation end moraine.

shaped clastic dykes strike across the surface slope. These are described in Appendix 2. The contact with the overlying outwash gravel is sharp and scoured.

Unit 6.

The outwash gravel consists of horizontally bedded sands and gravels that dip west by up to 15°. The gravel is well sorted, has numerous voids many of which contain illuvial mud. Beds of sand within the gravel are coarse, poorly sorted, massive and locally iron-cemented to the gravels.

The pebble fabric of the dip direction of the A/B plane of disc-shaped clasts suggests deposition by a current flowing toward 227°, obliquely away from the reconstructed ice margin (Fig. 6.58). The lithology of the gravel is very similar to that of the underlying massive diamicton.

Gently warped silt with numerous small dish-shaped structures at the contact with the overlying diamicton show the contact to have been loaded.

Unit 7.

The uppermost deposit in the sequence is a very coarse diamicton with boulders up to 1.7 m in diameter resting in a silty, partly stratified matrix. Numerous thin, gently warped silt stringers are draped over larger clasts, and some small flow noses are preserved. The matrix of the diamicton is a dull orange colour from iron-staining and has multiple iron pans. The lithology of the diamicton consists of a mixture of locally derived and erratic rocks, and there appears to be a decrease in the amount of Jurassic dolerite and Permian sediments toward the top of the section (Figs. 6.58).

The diamicton has a loosely packed structure but is relatively cohesive. The pebble fabric is weak and unimodal with the direction of maximum clustering toward 164°.

Interpretation.

Units 3 and 5 are essentially identical. Because they contain lenses of stratified material and have weak pebble fabrics unrelated to ice flow direction, they are thought to have been deposited as sediment flows in a supraglacial depositional environment.

The highly consolidated nature, strong fissile structure and strong unimodal pebble fabric of unit 4 suggest it is a lodgement till. Because lodgement tills are always the basal sediments of an advance (Boulton 1976), the till is evidence that the glacier advanced beyond this site during the Blackwood advance.

The outwash gravel overlying the massive diamicton of unit 5 was deposited by a stream that probably flowed across the ice front from northeast to southwest. The overlying diamicton appears to have been deposited directly on the outwash surface.

The presence of large sub-angular erratic boulders up to 1.7 m in diameter, low compaction, pebble fabric and the washed character of many parts of unit 7 suggest that it was deposited by mass movement. Such movement probably included several small sediment flows and slumping from a supraglacial till on a melting ice surface. This is supported by the surface topography of the arcuate moraine ridge containing the supraglacial till which appears to have accumulated as a latero-frontal apron of debris adjacent to a melting ice edge.

The reconstructed sequence of events that produced these sediments is:

1. damming of the lower King Valley during the retreat of the King Glacier from the maximum position of the King advance;
2. formation of the thick deposit of laminated silts as bottom sediments in a large nonglacial or proglacial lake;

3. advance of the King Glacier, erosion of parts of the upper sequence of lake sediments and deposition of massive sediment flows interbedded with outwash gravels at the front of the glacier;
4. deposition of lodgement till during the advance of the glacier beyond the section that represents the Blackwood advance;
5. retreat of the glacier to a position near the section deposition of outwash gravels by a proglacial stream, followed by the formation of sediment flows, slump deposits and laminated silts that were formed at a relatively stable ice front. These sediments formed an arcuate latero-frontal moraine ridge;
6. post depositional deformation of the silts is associated with the formation of tension cracks and landslides.

The Governor River section.

In the lower King Valley, near the confluence of the King and Governor Rivers at G.R. 881305, a section exposed in the right bank of the Governor river shows a series of outwash gravels and tills separated by major unconformities (Fig. 6.59). The unconformities represented are between Thureau Formation tills and King Formation outwash gravels, and between King and Blackwood Formation outwash gravels.

Unit 1.

The lowermost sediments consist of a highly weathered diamicton that rests on and plugs tunnels in weathered Ordovician limestone as a 1.2 m-wide clastic dyke. The dyke appears to be an filled solution tunnel similar to that seen in Fish Formation outwash gravels (Fig. 6.40) and described in Appendix 2. The diamicton is highly weathered and weathering rinds on Jurassic dolerite clasts are up to 120 mm in thickness.

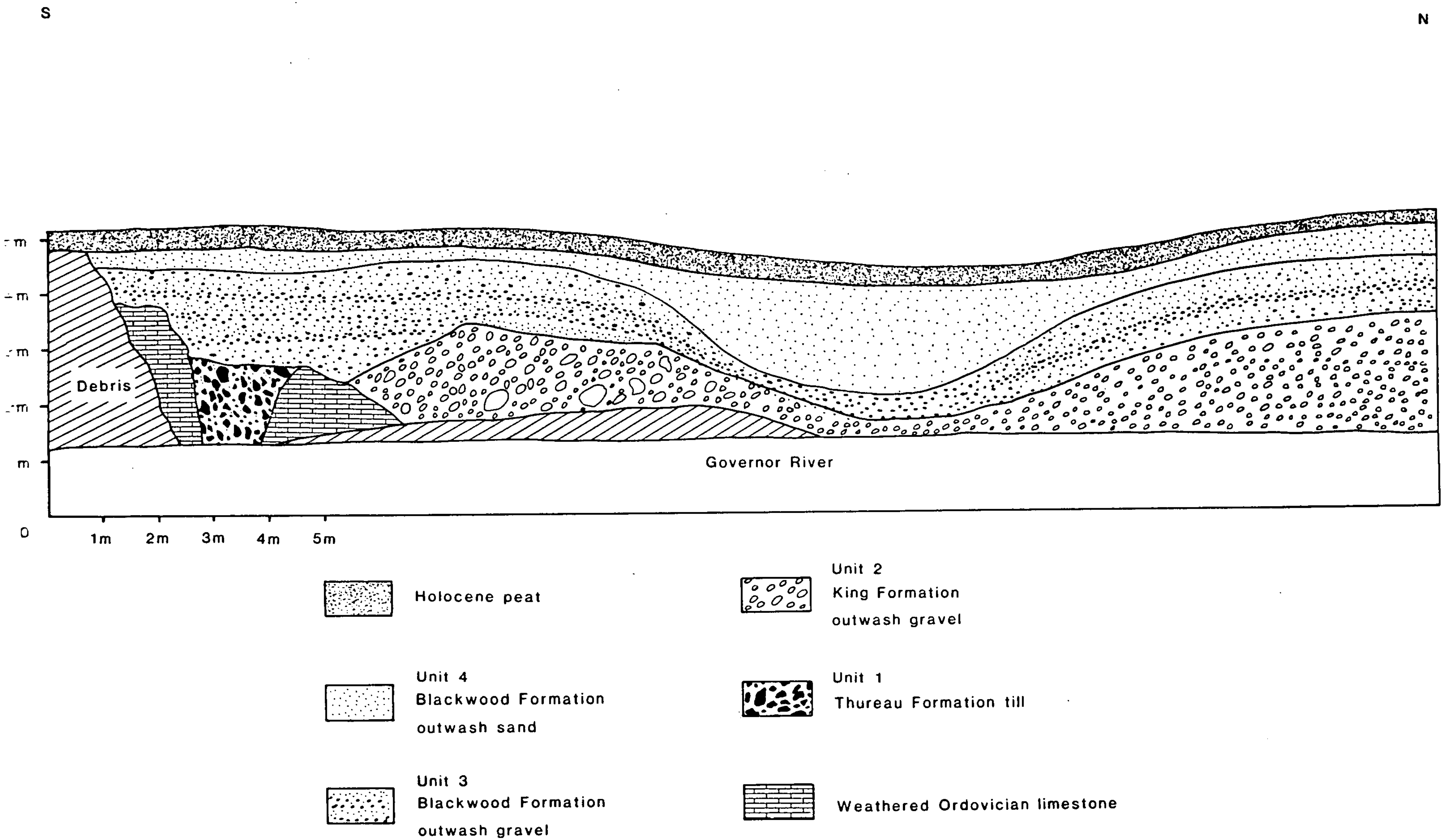


Fig. 6.59. Section through a Blackwood Formation terrace in the lower King Valley.

Unit 2.

The weathering rinds on dolerite clasts in the overlying bouldery gravel clasts have mean values of 9 mm which suggests a considerable period of time separates the deposition of the gravel and the underlying diamicton. However, it is not clear whether the gravel is part of the Governor or King formations, both of which are known to crop out within 500 m of the section. It is described as King Formation here largely on the basis of the proximity and altitude of known King Formation sediments.

The gravel of unit 2 is clast-supported and contains boulders of Ordovician conglomerate and Jurassic dolerite up to 1.9 m in diameter. Large logs buried by the gravel at the northern end of the section (Fig. 6.59) were dated at 35,200 $^{+800}_{-700}$ yrs. B.P SUA 2488. Because the wood was penetrated with humic acid throughout, it must be assumed that all old organic materials deposited close to river level at this site are likely to have been contaminated with modern carbon. Thus, the ^{14}C date should be regarded as a minimum age. The contact of this gravel with the overlying gravel is sharp and scoured (Fig. 6.59).

Unit 3.

Gravel overlying the coarse gravels of the King Formation are considerably less weathered and have weathering rinds with mean values of 3 mm. The poorly defined, steeply-dipping beds consist of moderately well sorted, imbricated gravels that probably accumulated on the face of a longitudinal bar. Again, the correlation of the gravel with the Blackwood Formation is uncertain and is based largely on the proximity of other Blackwood Formation gravels. These appear to form part of the same terrace that is inset into proximal outwash gravels of the King Formation. The contact between the gravel and the sand is sharp but there is no reason to believe that a major period of time separates their deposition.

Unit 4.

Up to 2 m of massive silty sands overlie the gravel and have a geometry that suggests they form part of a channel fill. This is supported by the surface geomorphology which is an abandoned channel. The channel can be traced upstream past the Regency sections and curves around in the direction of the lower King River Bridge. The sand is overlain by up to 0.6 m of Holocene peat.

Interpretation.

This section is similar to those at Baxter Rivulet and the Regency sections because it has a complex depositional history related to the successive reworking of sediments. The highly weathered till that forms a sedimentary dyke is probably a correlate of the Thureau Formation. Till of this age is known to rest on weathered Ordovician limestone in this area. The very coarse outwash gravels of the Governor or King formations almost certainly represent lags formed by erosion and redeposition of Thureau Formation till by meltwater streams. Outwash gravel overlying the lag gravel probably resulted from further reworking associated with outwash streams of the Blackwood advance. Absence of terraces below this level of reworking suggests that apart from minor areas of Holocene reworking, sediments in the lower King Valley have not been substantially eroded since the Middle Pleistocene Blackwood advance.

The change from outwash gravel to sand in the Blackwood Formation probably represents a facies change during glacial retreat with an ice-distal facies succeeding a proximal environment.

The Linda Creek sections.

At the mouth of Linda Creek there is a series of sections through glacial sediments at the same altitude as the type section of the Blackwood Formation. These sediments were deposited as ice-contact terraces and lacustrine deposits as the King Glacier blocked Linda Creek. They record a series of ice-marginal lacustrine environments dominated by high energy subaqueous debris flows.

The first of these sections is at G.R. 872409 and shows a series of interbedded gravels, sands and laminated silts that dip east into the King Valley at up to 28° (Fig. 6.60). The sediments record a delta deposited in a small ice-marginal lake.

Description.

Unit 1.

The sediments are described from west to east. The lowermost sediments consists of 1.2 m of gravelly sand that dip at 25° and grade into 1 m of massive sandy gravel. The contact with the overlying pebbly sand is sharp and appears to be scoured.

Unit 2.

The overlying pebbly sand grades upwards into finely laminated silts, sandy silts and silty sands, which are overlain by a thin gravel bed. Small pebbles of up to 30 mm diameter are dispersed in a massive sandy matrix. Clasts of Jurassic dolerite have weathering rinds of up to 12 mm in thickness. The sand is diffusely iron stained and locally bleached to a light grey colour. Sand grades upwards into finely laminated silts, sandy silts and silty sand in the upper 1 m. There is a sharp, eroded and loaded contact with the overlying gravel.

Unit 3.

The gravel is poorly sorted and has a maximum particle size of 100 mm. It is diffusely iron stained and cemented to the underlying silty sand. The gravel has a large boudinage structure (Figs. 6.60 and 6.61) which appears to be due to coherent slumping of both the gravel and overlying laminated silt. The laminated silt is intruded from below by the gravel in several places (Fig. 6.62) The underlying silts are sheared at the contact of the gravel and show small scale boudinage structures, convolute lamination and chaotic deformation.

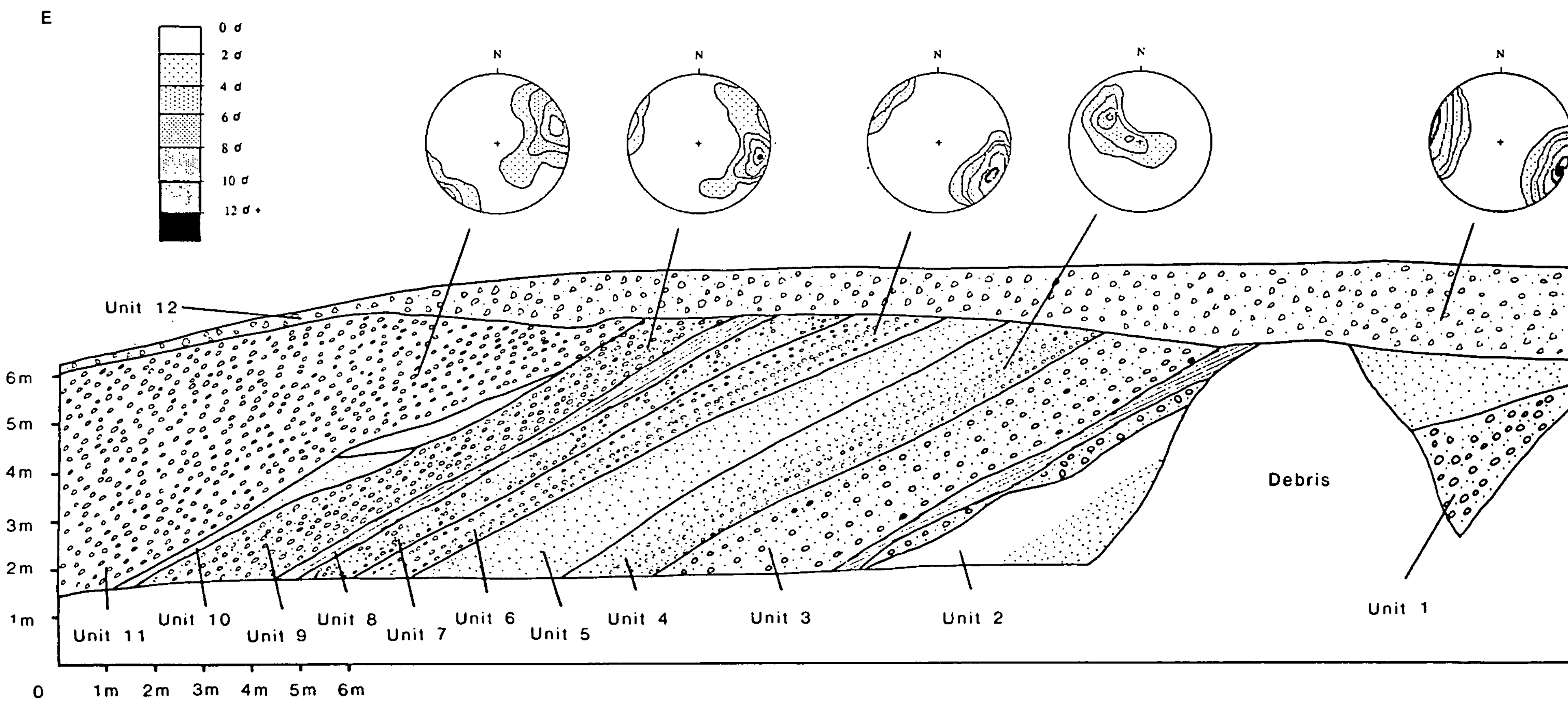


Fig. 6.60. Sediments exposed at the mouth of the Linda Valley.

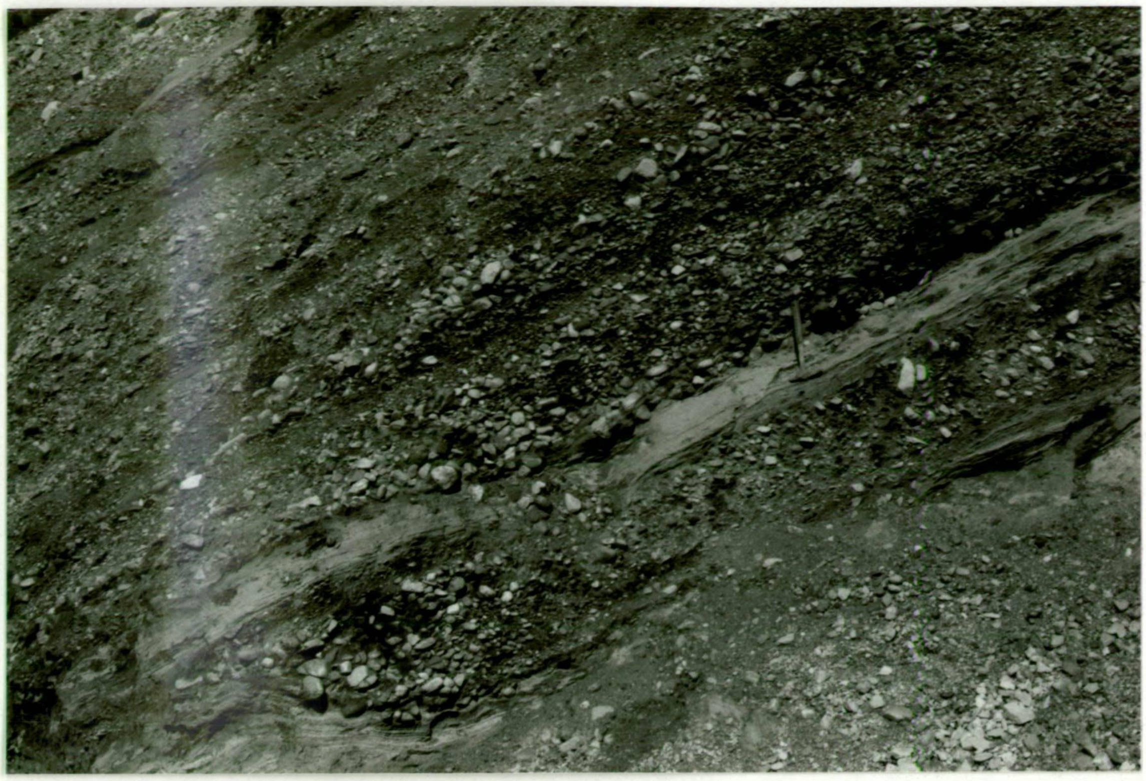


Fig. 6.61. Large boudinage structure in gravel overlain and underlain by laminated sandy silt.

Fig. 6.62. Steeply dipping gravels and laminated sandy silts exposed at the mouth of the Linda Valley. Note the thrust in the silt bed in the central part of the photograph.

Coherent slumping has caused dislocation of the gravel which is pinched into two lenses by the overlying and underlying silts. Small, sand-filled tension gashes in the upper surface of the overlying laminated silts appear to have formed at the same time as the boudinage structure.

Unit 4.

The contact between the sand and overlying gravel is gradational over 0.3 m. The 2.2 m of gravel consists of several small-scale upward-fining cycles. All the beds dip at near 30° to the east. The lithology of the pebbles is predominantly of locally derived siliceous rocks with rare erratic clasts of Jurassic dolerite (Fig. 6.60). This is typical for all the underlying sediments in which the proportion of erratic clasts does not increase beyond 10%. The lower 1.2 m of gravel is black and partly cemented and stained with iron and humic acid precipitate. The pebble fabric of the gravel reflects the slope it was deposited on (Fig. 6.60). The contact with the overlying sand is sharp, and is probably conformable.

Unit 5.

The overlying sediment consists of 1.7 m of massive sand which grades up into laminated silty sands. Numerous isolated pebbles up to 100 mm in diameter and frequent thin gravel lenses are common in the massive sand. Several of the small gravel lenses have flow structures and shear banding at their lower margins suggesting that they were deposited as grain flows. Laminated silts at the top of the unit are a pale grey colour. The underlying massive sands are iron-stained and have multiple iron pans. The contact with the overlying gravel is marked by gentle warping and load structures.

Unit 6.

The overlying deposit is a well sorted, steeply dipping, bedded gravel with a maximum particle size of 100 mm. The pebble fabric of the dip direction of A/B planes of disc-shaped clasts reflects the eastward dip of the bedding planes. The contact with the overlying gravelly sand is gradational over 50 mm and is marked by a thin bed of medium sand.

Unit 7.

The overlying sand consists of 0.7 m of poorly sorted sand with numerous isolated pebbles up to 60 mm in diameter dispersed throughout. The contact with the overlying silt is sharp and appears to be draped over a depositional surface.

Unit 8.

The overlying layer of silt consists of 0.8 m of pale green laminated sandy silt with numerous dropstones dispersed throughout the matrix. It grades upwards into a grey laminated silt without dropstones. The contact with the overlying sand is sharp. The top 30 mm of the silt is highly deformed and has been churned up into the matrix of the gravel.

Unit 9.

The overlying sediment consists of 1.3 m of poorly sorted, silty sand with numerous pebbles up to 100 mm. in diameter. The upper and lower contacts of this sediment body converge upslope and it appears to have a lens-shaped geometry. The pebble fabric reflects the depositional slope which dips at 32° to the east (Fig. 6.60). The lithology suggests a predominantly local source with rare clasts of erratic rocks. The sand is locally stained black by humic acids and has numerous small iron pans. The contact with the overlying sandy silt is sharp and scoured.

Unit 10.

The overlying 300 mm to 600 mm-thick bed of sandy silt contains a thin gravel lens (Fig. 6.60). It thins upslope, so that the overlying and underlying sediments are indistinguishable from each other. A low-angle thrust fault suggests that the upslope part of the silts slid down and over itself (Fig. 6.60). The sandy silt fines upward into laminated clayey silts. The contact with the overlying gravel is sharp and appears to represent a scour surface.

Unit 11.

The overlying 3.5 m of well sorted, clast-supported gravel which dips to the east at 28° is massive, diffusely iron stained and has multiple iron pans. The pebble fabric of the gravel reflects the eastward dip of the bed. The lithology is dominated by locally derived siliceous rocks.

Unit 12.

The diamicton which unconformably overlies all the dipping sediments is clast supported, crudely bedded and has a maximum particle size of 300 mm, which is much larger than in any of the steeply dipping sediments. Pebble fabric of the dip direction of the A/B planes of disc shaped clasts of the bedded part of the gravel suggests that it was deposited by a current from east to west. This indicates deposition up the Linda Valley against the slope of the older dipping sediments. High angles of dip such as these are not typical of braided river depositional environments which generally have pebble fabrics that dip upstream at between 26 and 32° (Rust 1975). The high dip angle may be from disturbance by mass movement of the deposit away from the adjacent rock walls.

Interpretation.

Because of the steepness of the depositional slope, the pebble fabric of individual beds, the presence of multiple occurrences of laminated silts with dropstones, and zones of slip and shear, the section is interpreted as the foresets of a prograding delta.

The delta prograded from west to east, that is toward the ice dam and not from the ice margin. The source of debris is unclear but the high altitude of the deposits relative to the valley floor suggests that the source was an ice mass that remained in the Linda Valley long after the initial glacial retreat.

The predominant mode of transport on delta foreset slopes is by mass flow and avalanching (Potsma and Roep 1985), and is dominated by bedded gravels and sands often with multiple scour surfaces (Cohen 1979). The multiple occurrences of laminated silts seen in this section are not typical of delta foreset environments, and are interpreted as being draped over sliding and avalanching gravels during quiescent periods. Dropstones within the laminated sediments suggests that there was floating ice on the surface of the lake during deposition of the laminated sediments.

The delta is unlike other glaciolacustrine deltas described in the literature that are largely composed of bedded sands and gravels. The laminated silts usually form on flat beds as bottomset sediments, not on steep foreset slopes as they have in this section. I am unaware of any analogue modern or ancient, for these laminated sediments which appear to be a large scale version of drape lamination as described by Gustavson *et al.* (1975). That is, they formed as silt was deposited from suspension and were draped over pre-existing topography and bedforms.

An alternative explanation is that the whole sequence of deposits accumulated on ice and was lowered into this position during melt of the supporting ice. However, because the pebble fabric of the gravels suggests deposition on a steep slope and because the sediments lack the intense deformation that would accompany lowering, the sediments are thought to have accumulated on a delta slope.

The Linda Creek section 2.

About 100 m west of the section described above at G.R. 870409, a large section 300 m long and 15 m deep exposes a sequence of deposits that have a similar geometry to the section described above. Although most of the section is covered by moss, occasional clearings show diamictons interbedded with steeply dipping silts. The sediments appear to have accumulated on a slope that has the same geometry as the delta deposits described above.

The eastern edge of the section shows a sequence of laminated silts underlain by gravels and overlain by lodgement tills and slope deposits (Fig. 6.63). The laminated silts lie below the series of subaqueous sediment flows that are poorly exposed in the western extension of the section.

Unit 1.

The lowermost unit rests on angular, fractured Ordovician conglomerate that has been plucked and eroded by glacier ice. It consists of a coarse diamicton with clasts up to 2 m in diameter. The pebble fabric of the diamicton shows a weak girdle pattern that does not appear to be directly related to the ice flow direction.

Unit 2.

The diamicton is overlain by 0.5 m of yellow laminated silt. The contact between the diamicton and the silt is scoured. The silt is draped over the boulders that protrude from the surface of the diamicton.

Unit 3.

The gravel that rests on the silt is well sorted and has numerous voids that are partly illuviated with silt. The pebble fabric of the gravel, measured from the dip direction of the A/B planes is strong, unimodal and suggests it was deposited by a stream flowing from west to east (Fig. 6.63).

Unit 4.

The gravel is overlain by about 3 m of pale yellow very finely laminated silt. The upper 2 m of the silt is deformed and the intensity of the deformation increases towards the contact with the overlying diamicton. The deformation structures include abundant small reverse faults boudinage structures, large-scale warping and thrust faults. The strikes of the reverse faults are not consistent but most have a steep dip (Fig. 6.63). The large warping and the truncation of the silts suggests that they have been dragged up and sheared off by overriding ice.

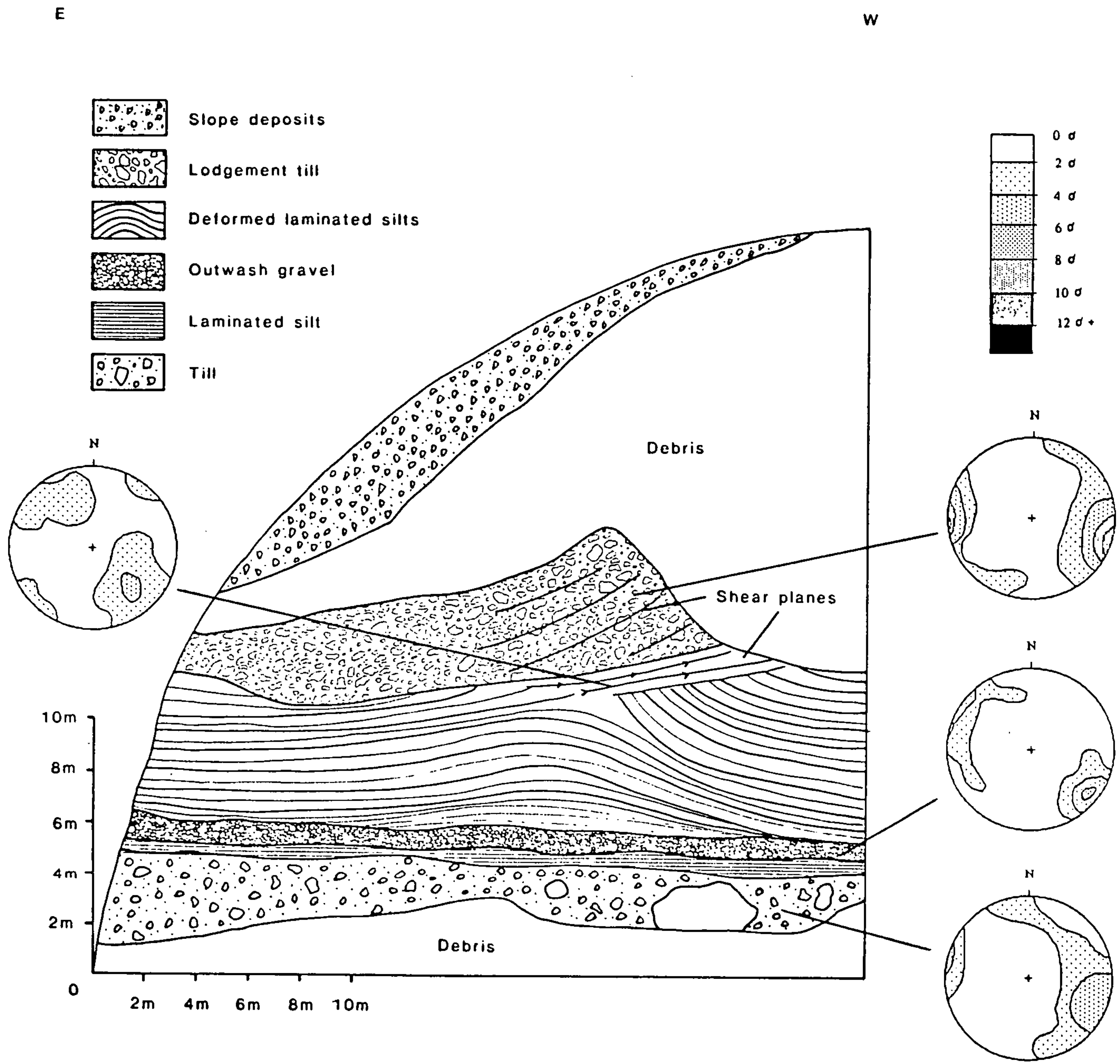


Fig. 6.63. Sediments exposed 100 m west of those shown by Fig. 6.60.

Unit 5.

The deformed laminated silt is overlain by a highly consolidated diamicton that consists of pebbles in a silty matrix. The diamicton has a fissile structure and multiple arcuate shear planes that run through it from east to west. The pebble fabric of the diamicton is unimodal and appears to be closely related to the direction of ice movement which was from east to west, up the Linda Valley (Fig. 6.63).

Interpretation.

The sequence of sediments has clearly been deposited in an ice-marginal lake that was overridden by ice.

The origin of the lower diamicton is not clear although from its location and pebble fabric it seems likely that it is a deformed lodgement till. The overlying outwash gravel was deposited by a high energy stream that flowed on the surface prior to the formation of the lake in which the silts were deposited. The absence of dropstones and sediment flows in the silt suggests that they were deposited during a relatively quiescent phase, possibly during the onset of an advance when ice formed a dam across the mouth of Linda Creek. The deformation structures in the silt are all post depositional and formed when glacier ice advanced over the lake deposits, deforming the silt and depositing a bed of lodgement till.

The remainder of the large section that lies to the west and above the sediments is not clearly exposed. However, the limited exposure of the sediments suggests they consist of large subaqueous sediment flows. The geometry of the sediments suggests that they accumulated in an ice-marginal lake on a steep delta slope. The Linda Creek section 1 appears to contain the final sediment sequence to have been deposited on a foreset delta slope in another lake that formed after ice retreat.

Figure 6.64 is a reconstruction of the environment of deposition, an ice marginal lake with a delta prograding into it from Linda Creek.

The sequence records two small ice advances, both of which were followed by periods of deposition in a relatively stable ice marginal lake. It is not known how the events recorded relate to the advances recorded in the King Valley but they may be related to the Blackwood and Bull advances.

Valley Creek sections.

Two sections near Valley Creek at G.R. 932383, near the Nelson River, record an ice advance over a proglacial outwash surface. They are the only two sections in this area that record the terminal sediments of the distributary ice lobe that flowed into the Nelson River during the Henty Glaciation. Several end moraines at Valley Creek appear to record a fluctuating glacier terminus. About 2.2 km further south near the Nelson River cave (G.R. 918366) a small section of glacial sediments appear to record the maximum extent of ice in this area during the Henty Glaciation.

Description.

The first section records outwash sediments that crop out just inside a moraine (Fig. 6.65).

Unit 1.
The lowermost sediment consists of 1.9 m of laminated sand that grades upwards into massive sand and rests on weathered Siluro-Devonian siltstone. The sand is poorly sorted and silty at the bottom. Sorting improves up the section which mainly consists of well sorted medium sand. The contact with the overlying gravel is sharp and scoured.

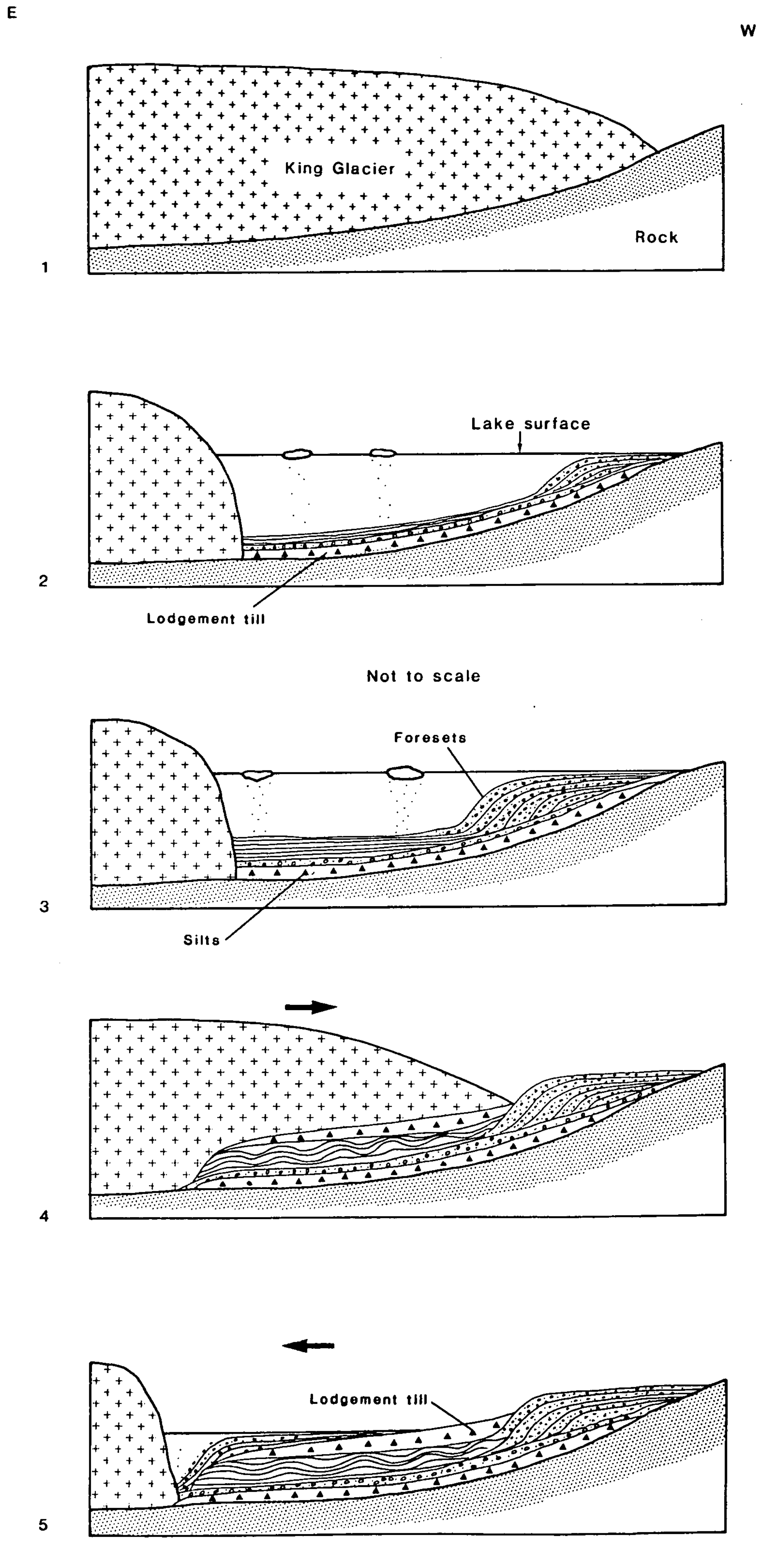


Fig. 6.64. Reconstruction of the Linda Creek depositional environment.

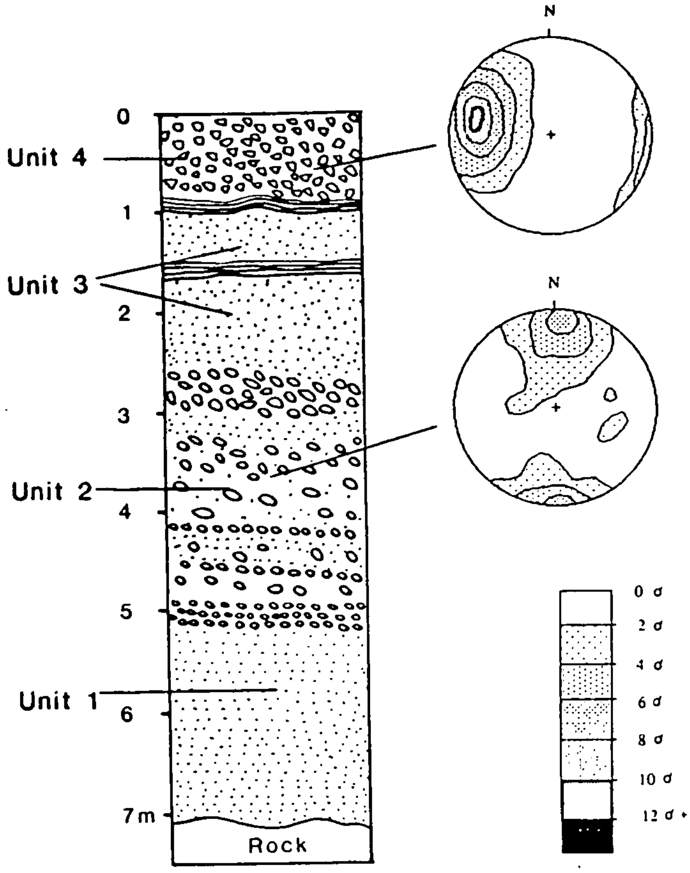


Fig. 6.65. Valley Creek section 1.

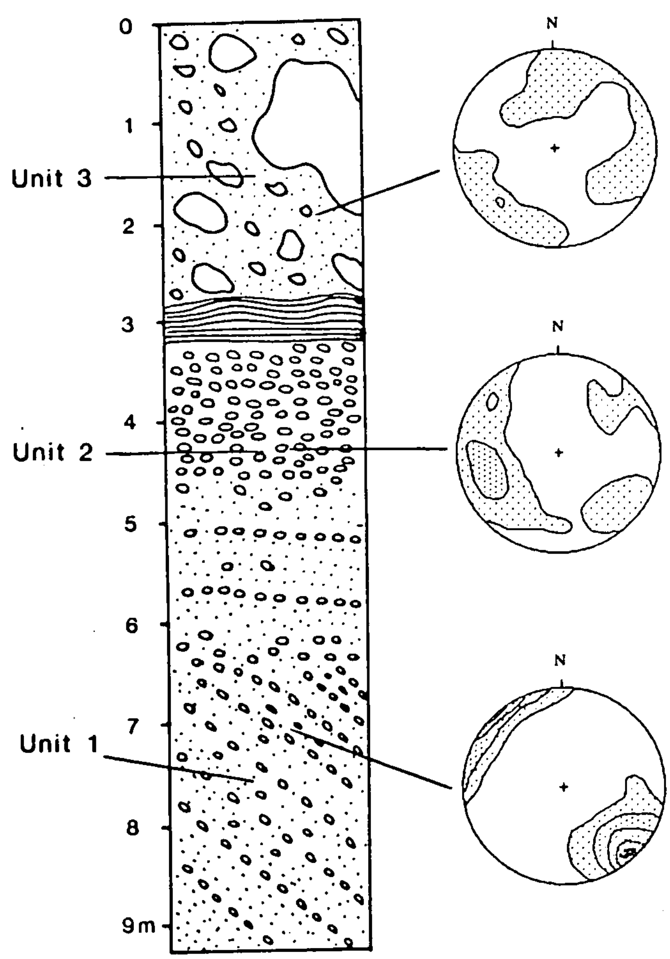


Fig. 6.66. Valley Creek section 2.

Unit 2.

The overlying gravel consists of a 200 mm-thick bed of well sorted gravel, 1.2 m of sandy gravel, 0.5 m of coarse, well sorted, gravel, 200 mm of well sorted massive medium sand and 0.5 m of coarse well sorted gravel. The lithology of the gravel consists of a mixture of erratic, West Coast Range and Eldon Range rocks.

Unit 3.

The overlying sand consists of well sorted, massive medium sand with two thin layers of very finely laminated silt near the top. The contact with the overlying gravel is sharp and erosional.

Unit 4.

The overlying gravel is poorly sorted and consists entirely of locally derived Siluro-Devonian quartzite from the adjacent steep slopes. All clasts are angular to subangular and rest in a matrix of fine sand which reflects the particle size of the quartzite from which it was derived.

Valley Creek section 2.

The second section occurs 30 m from the first but there are significant changes in the nature of the sediments.

Unit 1.

The lowermost sediments consist of 2.5 m of moderately well sorted gravel with beds that dip southeast at 20°. The pebble fabric measured as the dip direction of A/B plane dips at 18° toward 140° and reflects the slope of the bedding plane on which the gravel accumulated (Fig. 6.66).

Unit 2

The steeply dipping gravel grades into 3 m of horizontally bedded outwash gravel that has a weak pebble fabric (Fig. 6.66). The gravel is poorly sorted and has a maximum particle size of 60 mm.

Unit 3

The outwash gravel is overlain by 0.4 m of finely laminated silt that is iron stained and cemented. The silt is capped by massive bouldery diamicton that contains erratic Jurassic dolerite boulders up to 1.6 m in diameter. At the contact the silt is warped and has some small load structures which suggest the silt was plastic when the diamicton was deposited on it. The pebble fabric of the till is weak and unimodal (Fig. 6.66).

Interpretation.

Sediments of this section record an ice advance over a proglacial outwash surface. The lowermost gravel appears to have been deposited on the face of a longitudinal bar. Outwash gravel overlying the bar records an aggrading river bed, possibly associated with an increasing sediment flux from the ice advance. The massive till appears to be a dump moraine that is clearly visible on aerial photographs.

6.11 The Chamouni Formation.

The Chamouni Formation is limited in extent to the upper King Valley near the Lyell Highway. It consists of a series of terraces inset in older glacial sediments that can be traced 4 km downstream of the type section to a position near the Blackwood Formation terminal moraine (Map 1). Although most exposures of the terrace consist of bedded outwash gravels unconformably overlying laminated silt, an exposure near Comstock Creek shows that part of the terrace consists of lodgement till.

The type section.

The type section of the Chamouni Formation consists of a small 2.4 m-deep section of highly consolidated lodgement till on the upstream end of a roche moutonnee at G.R. 896432. The surface morphology of the outcrop is flat and appears to be part of a subglacial till sheet. The sheet forms part of an extensive terrace at 236 m altitude which is topographically above the lower part of the Dante outwash fan (Fig. 6.67).

Description.

The section consists of 2.4 m of highly consolidated dark grey silty till overlain by a thin soil. In the upper 1.6 m the structure of the till exhibits a sub horizontal fissility. The fissility dips to the north at between 5 and 10°, and appears to be due to shearing during or after deposition (Fig. 6.68). There is no textural variability within the diamicton which consists of subangular to rounded clasts of mixed lithology up to 150 mm in diameter.

Weathering rinds on Jurassic dolerite have a mean thickness of 1.53 mm and a standard deviation of 0.7 mm, which is very similar to that described by Kiernan (1983a) for the Dante Formation. The matrix is unweathered apart from some minor bleaching and iron-staining which is restricted

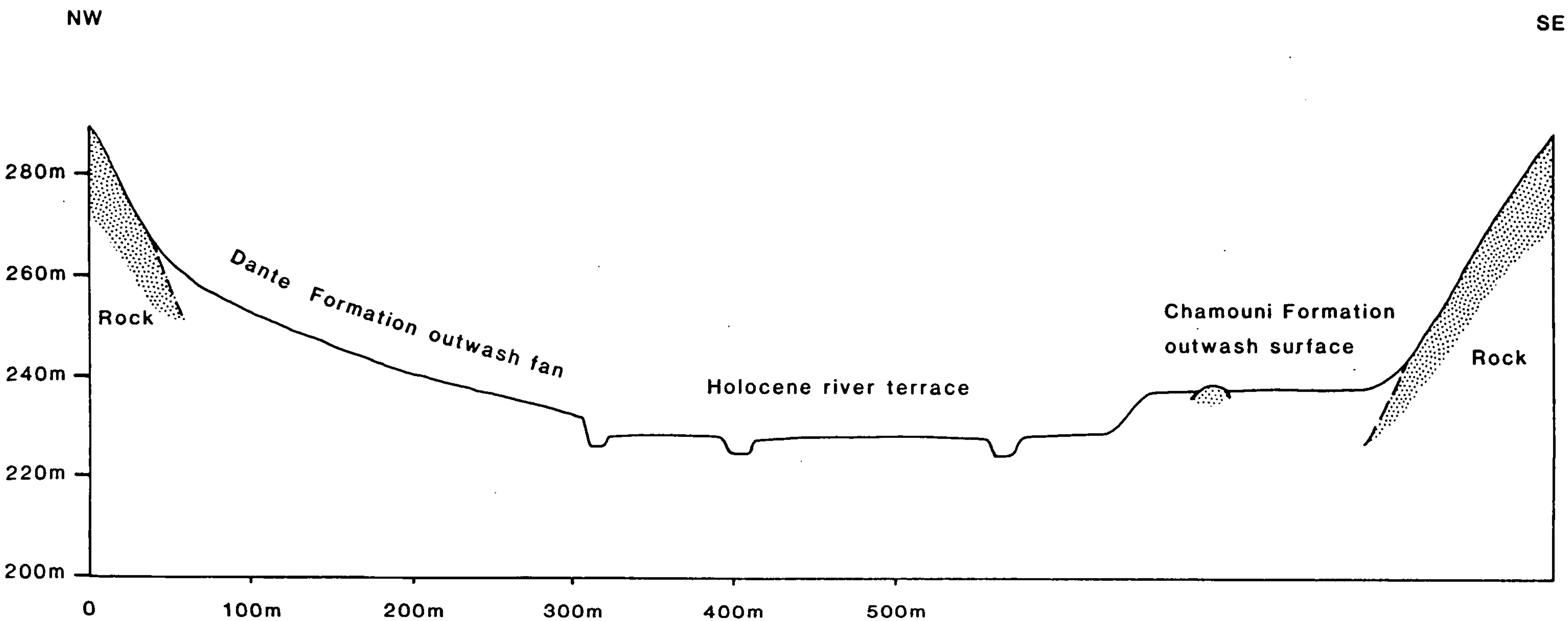


Fig. 6.67. The relationship between the aggradation surfaces of the Chamouni and Dante formations.

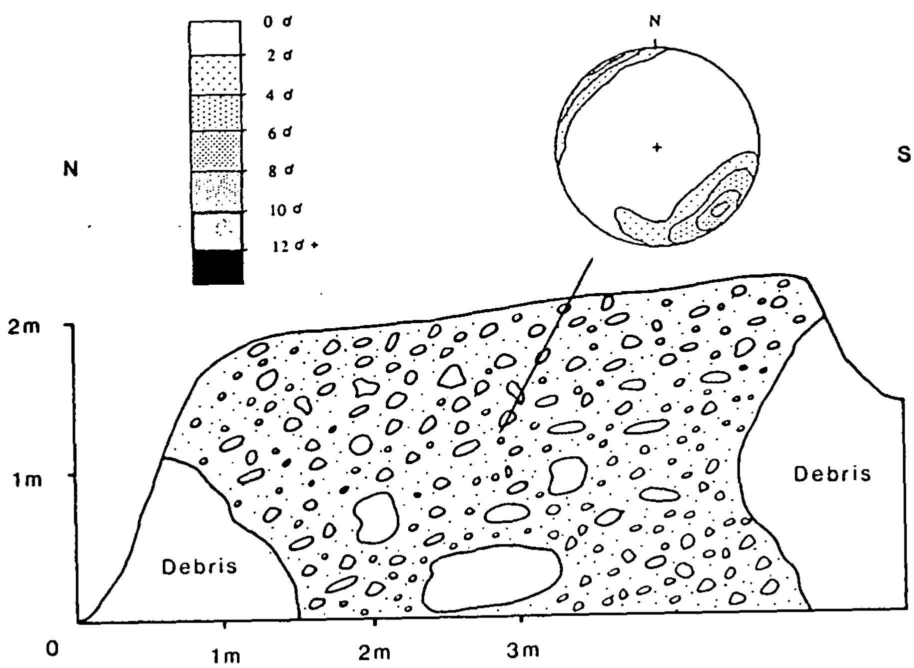


Fig. 6.68. The Chamouni Formation type section.

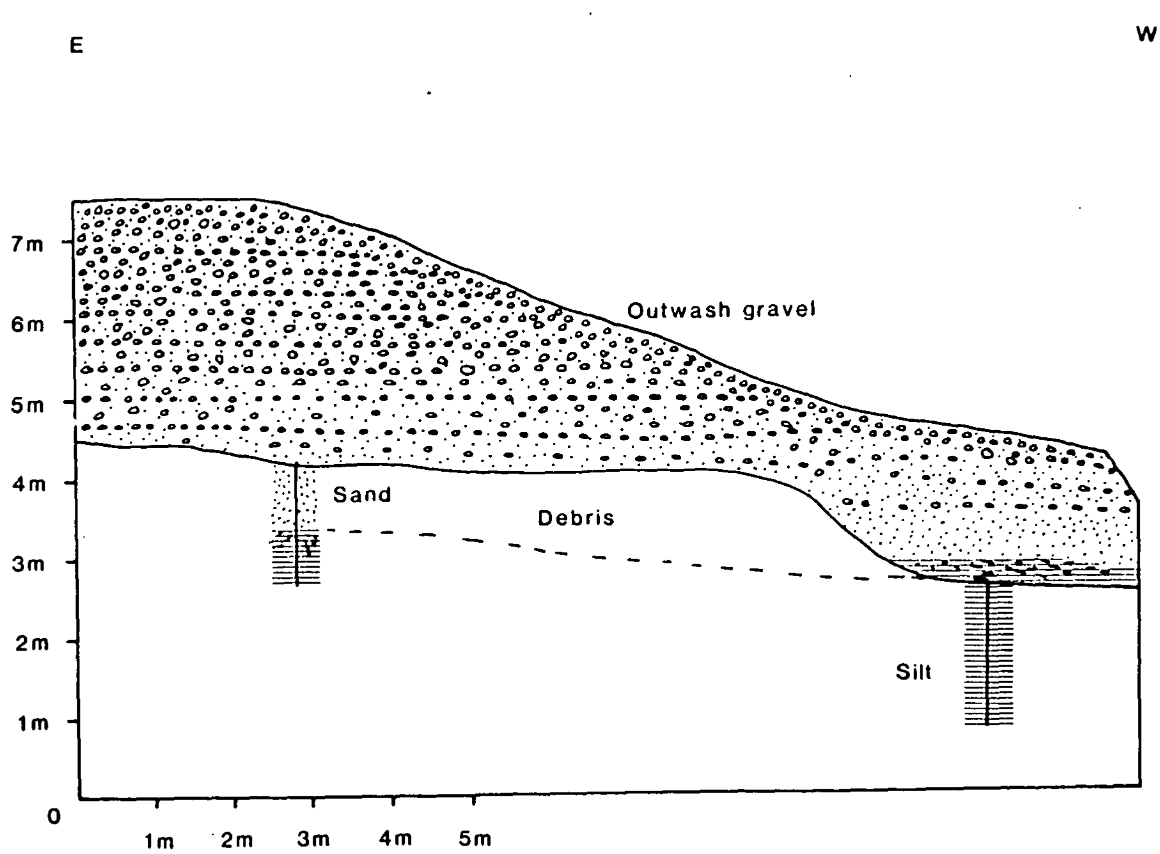


Fig. 6.69. Chamouni Formation sediments 800 m south of the type section.

to the top 200 mm of the soil profile.

The pebble fabric of the diamicton is strong and unimodal with the direction of maximum clustering (V_1) being toward 151° and dipping at 17.8° (Fig. 6.68). This is consistent with the reconstructed ice flow direction from striae on nearby roches moutonnees.

Although the thickness of the sediment is not known, an outcrop of Siluro-Devonian siltstone 60 m south of the section suggests that the till is not very thick and rests on bedrock. Although the surface of the terrace has been eroded by a number of shallow channels, the surface of the till shows no signs of a lag deposit and appears to be a depositional surface.

Interpretation.

The highly consolidated nature of the diamicton together with the fabric, fissility, texture and its location on the upstream side of a roche moutonnee, suggest that it is a lodgement till. The only apparent inconsistency in the interpretation is that the pebble fabric dips down glacier whereas other described pebble fabrics of lodgement tills dip up glacier (Boulton 1976). This may be due to the local stresses and flow deflection on the upstream side of a rock obstruction.

Further downstream, the terrace consists of several small roches moutonnees rising above the terrace surface, and numerous exposures of outwash gravel resting on organic-rich sands and silts.

King River Road section.

About 800 m south of the type section near the King River Road at G.R. 889427 an exposure in the Chamouni Formation terrace shows an upward coarsening sequence from organic silts to sand, coarse sand, pebbly sand and gravel (Fig. 6.69). Figures 6.67 and 6.70 shows the relationship between the Chamouni, Dante and Holocene terraces. An infinite radiocarbon date (interpreted as >48 ka) has been obtained from wood deposited near an unconformity between Holocene gravels and silts. The unconformity can be traced beneath the Chamouni Formation terrace (Fig. 6.70). The overlying outwash appears to record the onset of the Chamouni advance and the burial of laminated lake silts.

Description.

Although the depth of the dark grey silts is unknown, auger holes show that they are at least 4.5 m thick. The silts are very finely laminated, with individual laminae from less than 1 to 4 mm thick. Abundant small drifted wood fragments, all less than 2 mm in diameter, occur in the uppermost 4 m of the silt which smelt sulphurous and appeared to contain a high organic content. Five samples at various depths were processed to extract pollen but only five pollen grains of Compositae were observed in all the samples (G. van de Geer pers. comm. 1986). The silt grades upwards into massive medium silty sands and fine pebble gravels. The transitional sands show a sudden decrease in organic content at 3.3 m.

The overlying gravel is up to 2.8 m thick and consists of well sorted, rounded and horizontally bedded pebble-gravel with a coarse sand matrix. The lithology of the gravel consists of a wide variety of rocks but is dominated by locally derived rocks. The gravel is diffusely iron stained and isolated patches are iron cemented.

Interpretation.

The upward coarsening sequence seen at the first section shows that the gravels are conformable on the silts. The sequence clearly represents the onset of an ice advance. Weathering rinds on Jurassic dolerite in these gravels and on the till seen at the type section suggest they are of similar age to the Dante Formation and much younger than the Blackwood Formation.

King River Road section 2.

An excavation near the King River Road at G.R. 885422 shows low-lying Holocene gravels unconformably overlying the grey silts of the Chamouni Formation.

Description.

The lowermost deposits in the excavation consist of dark grey silts that are finely laminated and apparently identical to silts that crop out in the section described above. In this section the silts grade into sands (Fig. 6.70) but unlike the other sections the sands contain organic debris that includes drifted twigs and leaves. Pollen from this section clearly represents an alpine to subalpine flora which supports the interpretation that the transition from the silts to the outwash gravels represents the onset of a glacial advance (Table 4.3). The wood was dated at $48{,}700^{+2900}_{-2100}$ yrs. B.P.(SUA 2599) and is overlain by low-lying Holocene river gravels. Figure 6.71 demonstrates the relationship between the Holocene terrace and the Chamouni Formation terrace.

Interpretation.

Since the contact between the silts and the outwash gravel is conformable the clear implication is that the Chamouni Formation is considerably older than the Dante Formation. The age of the formation is discussed further in section 8.2.

Excavations in the southern most extension of the Chamouni Formation terrace, 3.2 km south of the section described above at G.R. 885404, show a similar pattern of outwash gravels overlying laminated silts. In these sections the silts grade upwards into massive sandy silts, sands and pebbly sand over 60 cm. At about 20 mm. below the contact with the overlying gravels the sand contained fragments of drifted wood and leaves. These were not dated but are believed to correlate with dated wood and leaves (Fig. 6.70) in a similar position upstream. The upper contact of the pebbly sand is sharp, scoured and dips south at 15°.

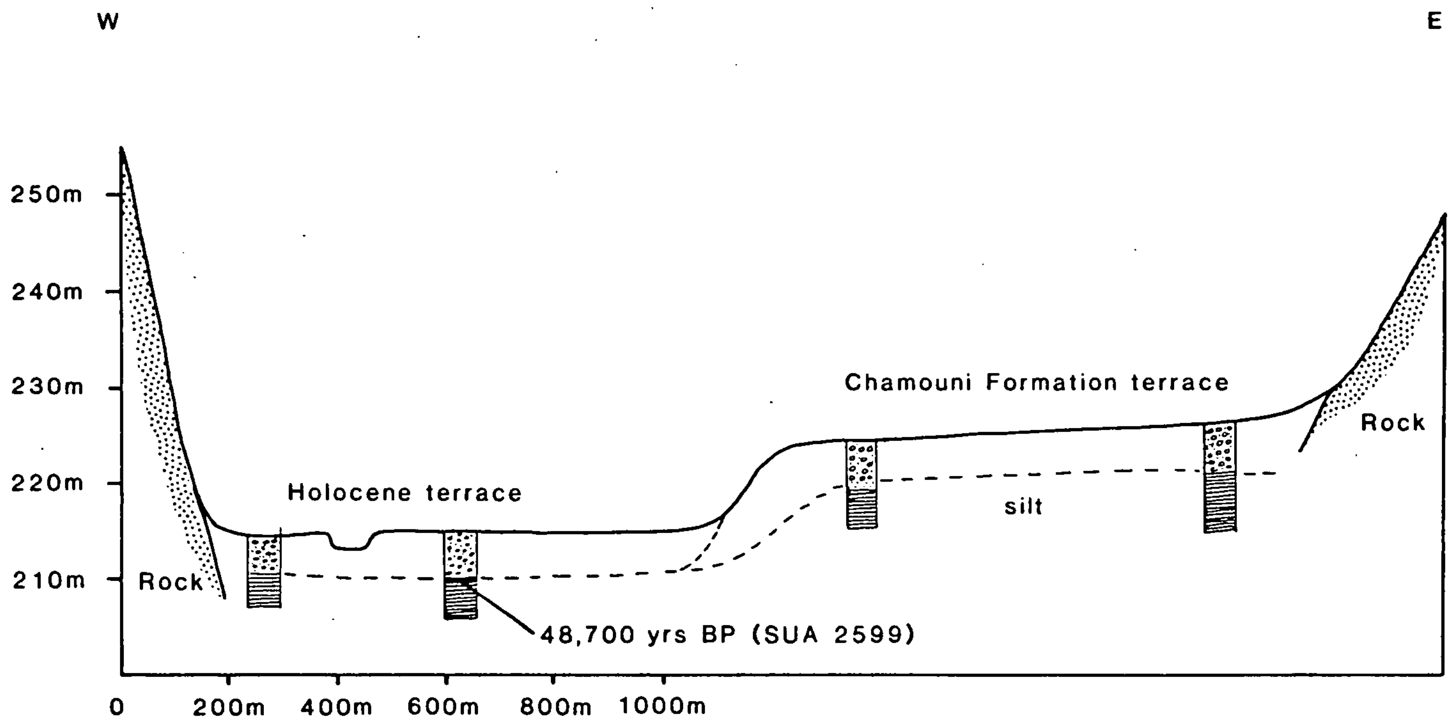


Fig. 6.70. Relationship between the Chamouni Formation and Holocene alluvial gravels.

6.12 The Dante Formation.

The type section.

The type section of the Dante Formation is exposed in the right bank of the King River as the distal part of an outwash fan from Lake Beatrice and Dante Rivulet. The section shows outwash gravels of unknown age overlain by organic silts, sands and outwash gravels of the Dante Formation (Fig. 6.71).

Description.

The gravels at the base of the section consist of well sorted, massive imbricated gravels with a mixed lithology. They are separated from the overlying gravels by a lens of organic sandy silt which forms part of a weakly developed alpine palaeosol.

The palaeosol is marked by vertical roots and aquatic plant stems, suggesting that the sandy silt was deposited in a wet, boggy environment, possibly with standing water. Pollen from the silt records an alpine herbfield-bog mosaic with a rich flora of herbs, sedges and low shrubs (Kiernan 1980, Gibson *et al.* 1987).

The uppermost gravels consist of 4.5 m of moderately well sorted horizontally bedded outwash gravel. Weathering rinds on Jurassic dolerite clasts have a mean thickness of 1.5 mm and a standard deviation of 0.27 mm. The lithology of the pebbles is a mixture of erratic, West Coast Range and locally derived rocks.

Interpretation.

Kiernan interpreted the section as demonstrating the formation of a soil during a period when the

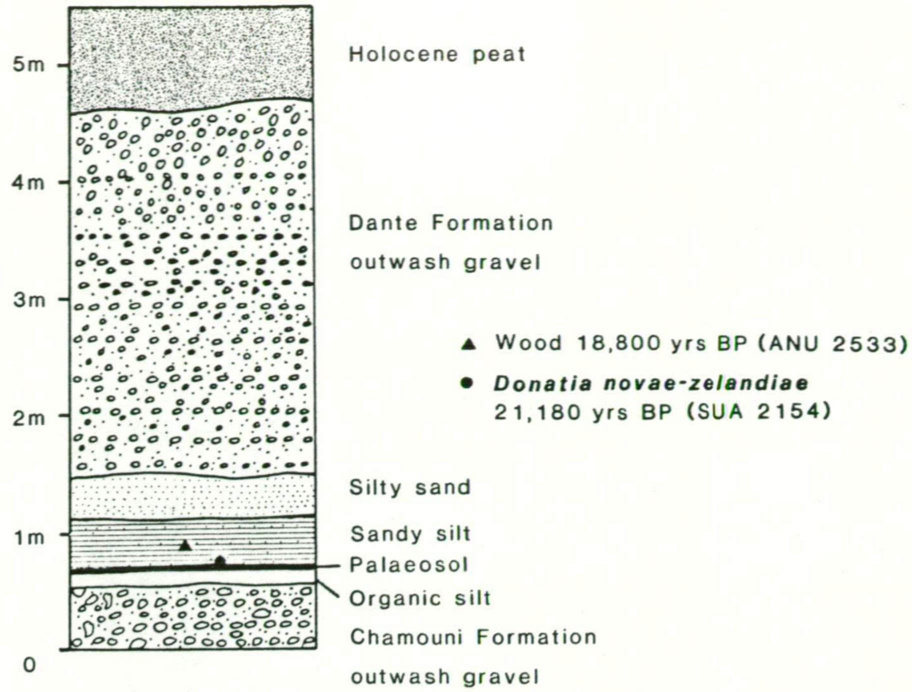


Fig. 6.71. The Dante Formation type section.

Fig. 6.72. Holocene river gravels. Wood taken from the base of this section was dated at 12,250 ± 90 yrs. B.P. (SUA 2415).

treeline remained below 230 m and that the grey sandy silt represents renewed glacial sedimentation after a break during which the palaeosol had formed (*ibid* p.35). Gibson *et al.* (1987) concluded that the wood dated at 18,800 ± 500 years BP (ANU 2533), a fossil bolster plant dated at 21,180 ± 370 years BP (SUA 2154) and twigs from the palaeosol dated at 20,100 ± 470 years BP (SUA 2155) imply that the overlying outwash gravels represent the late last glacial maximum.

An alternative interpretation of the lower gravels is that they represent an earlier advance of the last glaciation, and not part of the preceding glaciation as suggested by Kiernan (1980). They may be outwash gravel of the Chamouni Formation.

6.13 The Long Marsh Formation.

The type section of the Long Marsh Formation is at G.R. 882393 and consists of 3 m of coarse well sorted gravel overlain by 1.1 m of overbank silty sand (Fig. 6.72).

Description.

The gravel is moderately well sorted, crudely bedded and has a strong imbrication which suggests it was deposited by a current flowing from northwest to southeast. Below 1.6 m the gravel is diffusely iron stained and locally iron cemented with thin iron pans. The weathering rinds on Jurassic dolerite clasts have a mean thickness of 0.42 mm and a standard deviation of 0.16 mm.

At 3.5 m large logs up to 400 mm in diameter are buried in the gravel (Fig. 6.72). A core sample of a large log was ^{14}C dated at 12,250 ± 90 (SUA 2415). This date suggests that large trees were growing in the King Valley prior to 12 ka. B.P. and that the upper King Valley was ice free.

About 100 m northwest of the type section, excavations in the terrace show Holocene gravels resting unconformably on grey silt. The silt is the same sediment that is exposed below the Chamouni Formation terrace (Figs. 6.70).

Interpretation.

The King River level is now 3.5 m below the surface of the terrace and a meander is eroding the bank and depositing a point bar 2 m below the level of the terrace. Holocene terraces in the King Valley occur at two levels. One is within 1.5 m of base flow river level and is periodically inundated by surface waters. The other is 3.5 to 4 m above river level and is not subject to flooding.

Other Holocene terraces occur in the lower King Valley and in the Nelson River Valley (Map 1). They all show a similar pattern of coarse, well sorted gravel overlain by flood deposits of silty sands.

Other Holocene sediments.

Sediments of Holocene age other than those of alluvial origin are of relatively minor importance and are generally restricted to small areas.

The most common and most extensive of these are Holocene peats which form a blanket over most surfaces with less than 20° slopes. The thickness of the peat is usually around 600 mm but can be up to 1.2 m in low lying areas. A subsidence doline in the lower King Valley has accumulated 3.5 m of peat which overlies silts deposited during the last glaciation. Pollen analysis and dating of a 5.5 m core taken from the depression shows clearly that organic sedimentation in the form of peat began with climatic amelioration following the last glaciation. The bottom of the peat at 3.65 m shows the commencement of organic sedimentation and has been dated at 13,010 ± 130 yrs. B.P. (SUA 2723).

Sections at Newall Creek and on the northern end of the Thureau Hills record slope deposits of Holocene age within organic sediments. (Fig. 6.73 and 6.74), (van de Geer *et al.* 1988 in press). At the Thureau Hills section the layer of slope deposits occurs at 0.5 m depth just below wood dated at 2520 ± 80 yrs. B.P. (SUA 2597). Pollen analysis of the Pleistocene peat records a cold climate flora overlain by organic debris that records the successional establishment of temperate rainforest .

At Newall Creek 1.5 m of slope deposits separates a forest soil from organic silts which record a glacial advance and the successional establishment of temperate rainforest.

Together these two sites indicate that there was some slope instability within the middle part of the Holocene. Pollen analysis from other sites in southwestern Tasmania suggest the occurrence of a

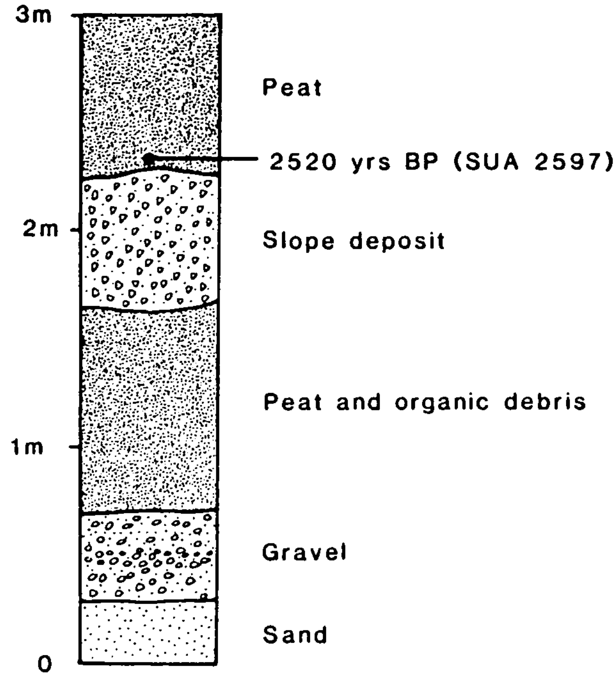


Fig. 6.73. Holocene organic and slope deposits adjacent to the Thureau Hills in the King Valley.

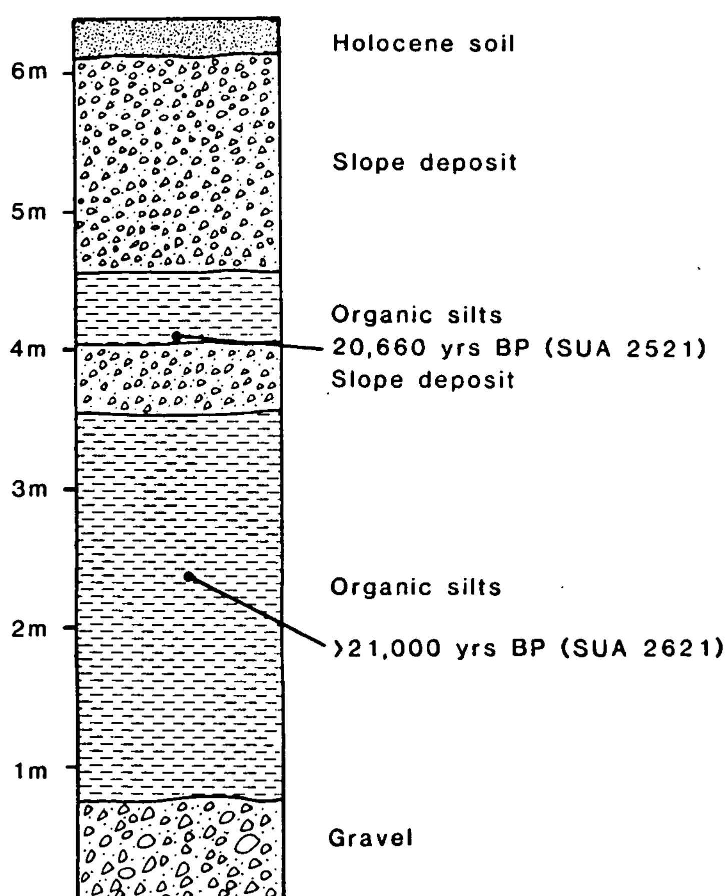


Fig. 6.74. Late Pleistocene to Holocene sediments exposed at Newall Creek.

mid Holocene moist period (Marksgraf *et al.* 1986), which lends some support to the patterns observed in Holocene sediments in the King Valley.

Thick slope deposits also occur on the northern face of Mt. Owen. Although the age of these deposits is not certain they are thought to be paraglacial sediments deposited during cold periods immediately after glaciation.

CHAPTER 7.

PATTERNS AND PROCESSES OF SEDIMENT GENESIS AND DISPERSAL.

7.1 Introduction.

The main controls over glacial deposition are glacier mass balance, thermal regime, bed configuration, the climate at or near the ice margin, the properties of the material being deposited and the amount and distribution of the debris within the glacier (Andrews 1975, Lawson 1979a). Although the basic controls of glacial depositional environments are the same, the relative importance of the processes in an area varies according to the combination of local environmental variables.

This chapter describes patterns that were observed in the sediments in the King Valley and discusses the controls on the nature and position of sediment deposition. Elements of the debris transport paths in the King Glacier are reconstructed and compared with what is known about glacial depositional systems of modern glaciers.

7.2 Processes of deposition in the King Valley.

The most abundant sediments in the King Valley are those produced by outwash streams and mass movements that have reworked tills. Primary tills deposited directly from glacier ice are relatively rare.

7.2.1 Primary tills.

Sediments of primary glacial origin or ortho-tills (Dreimanis 1983) are relatively uncommon in the King Valley. This applies specifically to tills of englacial and subglacial origin. Supraglacial morainic till as defined by Boulton and Paul (1976) is not regarded as a primary till because it is an assemblage of sediments formed by a wide variety of processes.

Lodgement and melt-out till are distinguishable by a number of characteristics that have been summarised by Boulton and Paul (1976), Shaw (1981), Haldrosen and Shaw (1982) and Lawson (1979a). Prominent among these criteria are those characteristics that have been inherited from glacial transport (eg. pebble fabric).

Melt-out till is very rare in the King Valley, only two occurrences are known. This is not unexpected since melt-out till does not usually survive in environments with large amounts of meltwater (Eyles 1979).

Lodgement till in the King Valley is known to occur only in five exposures. This till type has been recognised by its consolidated, fissile and sheared structure, and by its strong unimodal pebble fabric and dense, silty matrix. The relative scarcity of melt-out and lodgement is discussed below.

7.2.2 Secondary deposits.

Secondary or "resedimented" (Lawson 1979a) diamictons are those that have been deposited initially by primary glacial processes but have been reworked and their glacial character has been modified. Although secondary deposits have been recognised for some time (eg flow tills by Elson 1961 and Boulton 1968), it is only recently that their importance and variability of form has been described. Lawson (1979a, 1981a) estimated that up to 95% of the deposits forming at the margin of the Matanuska Glacier, a temperate glacier in southern Alaska, were formed by processes that reworked primary glacial deposits.

Secondary deposits are formed by a wide variety of processes that involve sediment flow, falls and slides (Lawson 1979a). Many secondary deposits have some stratification or sorting. In almost all cases fabric, particle size and sedimentary structures inherited from glacial transport are altered. The most common secondary deposits in the King Valley are sediment flows. These are found primarily in ice-terminal and supraglacial environments, frequently in association with lacustrine sediments and outwash gravels.

Deposits that have been interpreted as sediment flows in this thesis are recognised by their weak fabric strength, lack of relationship of pebble fabric to ice flow direction, identification of the palaeoslope on which the flow formed, flow structures and by intraformational blocks of deformed, laminated silt that have been eroded and transported by the flow.

In addition to sediment flows, other secondary deposits with highly variable characteristics are widespread throughout the King Valley. In many sections it is not possible to determine the mode of origin of these sediments with certainty. Although they frequently have the appearance and sedimentary characteristics of primary tills, they often have a weak pebble fabric that is unrelated to ice flow direction and structures related to non glacial procsses. Deposits of this type are usually seen as the basal sediments of sections and underlie, or are interbedded with, supraglacial sediments (eg unit 1, section 6.4 and units 3 and 5, section 6.10). The position of these sediments beneath supraglacial sediments suggests they are melt-out tills. However, they have few of the documented characteristics of melt-out tills such as a strong unimodal pebble fabric, parallel to the reconstructed ice flow direction, inclusions of subtill sediment, or intraformational blocks of stratified sediment (Shaw 1982; Haldrosen and Shaw 1982). Such deposits may have been formed in several ways. Some may be dump tills that were lowered or fell directly from the surface of the glacier, others may be deformed melt-out till, or proximal outwash gravel deposited so close to the ice margin that there has been little sorting of the material.

Dump tills are those that are released from the ice surface by falling and sliding onto subglacial or supraglacial surfaces. They have been described as ice-slope colluvium by Lawson (1979a) and as supraglacial till, ice-contact screes and ice contact fans by Boulton and Eyles (1979).

Melt-out tills are commonly mobilised in situations with high meltwater production (Lawson 1979a). Such reworking often results in the loss of some of the finer grained sediment and the acquisition of new sedimentary characteristics. In the King Valley, the abundance of outwash deposits suggests that meltwater was abundant during most phases of ice melt and that many of the unidentified secondary deposits may be post-depositionally modified melt-out tills.

Ice proximal outwash gravel can be transported and deposited with little sorting and stratification at short distances from the debris source during floods. Deposits of this type can resemble tills because of their poor sorting and lack of structures that clearly indicate water transport. Although these secondary deposits have the appearance of primary tills, the weak pebble fabrics and sedimentary structures that indicate disturbance suggests they are of not primary origin. They cannot be identified easily as having specific origins because descriptions of the genesis of such sediments in modern environments contain few criteria that can be used to identify ancient analogues with certainty.

7.2.3 Outwash sediments.

Outwash sediments are the most commonly occurring deposits in the King Valley. This is probably because outwash streams were concentrated in the relatively narrow valley floor and have reworked many of the older glacigenic deposits.

The appearance of outwash gravel throughout the King Valley is remarkably uniform. It is poorly sorted, massive or crudely bedded and has an average particle size of about 100 mm. In most cases it was not possible to distinguish between outwash gravel and alluvial gravel. Even the post glacial gravel has essentially the same appearance as older outwash gravel because it has been derived from glacigenic deposits and transported only a short distance prior to redeposition.

The sediments in the King Valley occur in a number of successions and associations that have distinct geographic patterns. The dominant successions are described below.

7.3 Sedimentary successions

Figure 7.1 is a facies relationship diagram (FRD). It is used here to demonstrate the successions observed in sections where diamictons are interbedded with stratified sediments. Most of the facies have been defined elsewhere in the text. The group "other secondary deposits" includes all deposits that show signs of being modified since their deposition as till but do not have evidence of having been deposited by current activity or flow.

The FRD has been constructed using the method of Selley (1970) which converts the number of transitions from one facies to another into probabilities. The probabilities are compared to a random transition matrix to identify transitions that are more and less common than random. The heavy lines on Figure 7.1 indicate transitions that are much more common than random and the light solid lines indicate transitions that are slightly more common than random.

The FRD can be divided into three areas that appear to represent three subenvironments; supraglacial, proglacial ice marginal and paraglacial (Fig. 7.1).

In the supraglacial environment there are strong transitions from lodgement till to sediment flows, outwash gravel and other secondary deposits. The strong link between lodgement till, sediment flows and other secondary deposits may be due to the surface of lodgement till being reworked or burial by supraglacial sediments that are lowered onto the surface as the glacier melts.

The subaqueous ice-contact environment is dominated by lacustrine sediments, ice rafted diamictons, subaqueous secondary deposits and outwash gravel. There are strong transitions between outwash gravel and lodgement till that are due to fluctuation of glacier margins in ice-proximal proglacial areas.

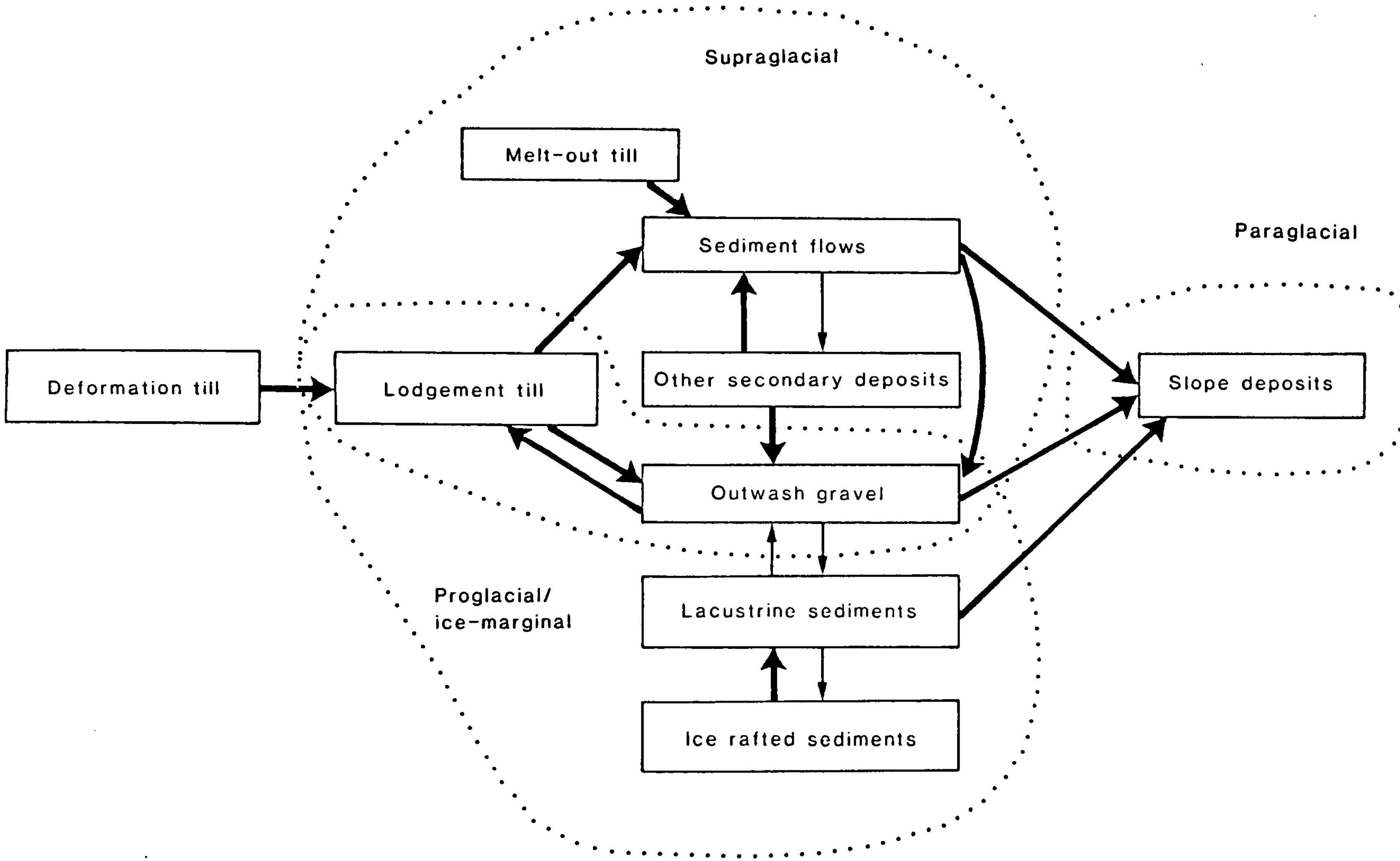


Fig. 7.1. Facies relationship diagram of sedimentary successions in ice marginal environments constructed using the method of Selley (1970). The heavy lines represent transitions that are much more common than random and the lighter lines represent transitions that are slightly more common than random.

Slope deposits from 1 to 5 m thick overlie most glacial sediments deposited in ice-marginal positions. The wide distribution of these deposits and their association with glacial deposits suggests they are paraglacial, i.e. their formation can be attributed to instability during and immediately subsequent to glaciation (Church and Rhyder 1972).

7.4 Patterns of sediment dispersal and sediment associations.

The sediments described in Chapter 4 occur in repetitive geographic locations and in similar associations with other sediments. This section summarises the major outwash, ice-contact and lacustrine sedimentary environments, and identifies the most important controls on the position and nature of the sedimentation.

7.4.1 Ice-contact sediments.

Three types of ice-contact depositional environment can be identified. These are supraglacial, frontal and subaqueous.

Supraglacial ice contact.

Supraglacial ice-contact sediments accumulate on melting ice and are usually deposited during glacier retreat. They are highly variable but generally consist of interbedded sediment flows, outwash sediments and lake silts. Sediments from this environment are characterised by a wide variety of syn- and post-depositional deformation structures. Sequences of deposits of this type were described from the Thureau, King and Blackwood Formations (sections 6.4, 6.8 and 6.10). The location and processes of deposition of sediments in this environment are directly related to the position of the glacier terminus, ice thickness and the debris content of the ice. Although the initial processes of deposition are independent of topography, often substantial disturbance occurs as the ice finally melts and the deposits are lowered on to the subglacial surface. Debris of this type appears to have been widespread in the King Valley and deposited during every advance/retreat cycle of the King Glacier. However, because the sediments were deposited during deglaciation, they were probably destroyed by subsequent outwash streams, glacial advances and alluvial reworking. Deposits that accumulate in supraglacial positions generally have a low preservation potential (Paul 1983) yet in the King Valley, sediments deposited in supraglacial positions are abundant. In the King Valley they probably form small remnants of what were formally extensive surpaglacial sediments and landforms.

The dominance of this sedimentary environment in the King Valley is probably related to the topography and glacial system which encouraged either very slow movement or stagnation of the ice. The slow-moving and stagnant ice probably formed because there are several major constrictions in the valley. When the glacier began to thin and ice velocity slowed, the constrictions would have tended to cut off ice supply to the lower parts of the glacier. The ice would then melt almost *in situ* and form large areas of stagnant ice with the potential for supraglacial deposition.

Ice frontal and latero-frontal sediments.

Ice frontal sediments are deposited at the glacier sides and terminus. They consist of a wide variety of sediments including dump tills, sediment flows, glacier contact fans and outwash gravel. The main areas of preservation of these deposits are the large end moraines such as that of the Blackwood Formation (section 6.10). The eastern part of this moraine appears to have accumulated as a dump moraine where flow tills and other secondary deposits overlie outwash gravel. The western part forms a ridge of outwash gravel that appears to have accumulated on or against the ice terminus. This situation is similar to that described by Boulton (1972b) who suggested that "moraines" formed in this way are quite common and are not true moraines, but are ice-contact ridges the position of which is controlled by the location and dynamics of the glacier margin.

Preservation potential of frontal and latero-frontal deposits is limited because they only accumulate in large quantities when the ice front position is stable for long periods of time.

They are very prone to erosion from outwash streams and successive ice advances.

Subaqueous ice-contact deposition.

Deposits in subaqueous ice-contact environments in the King Valley are dominated by laminated sediments interbedded with debris flows, delta deposits, ice-rafted diamictons, outwash gravels and rare subglacial sediments. Typical sequences of these sediments are the Thureau Formation sediments in the Linda Valley (section 6.4) and the Blackwood Formation sediments deposited at the mouth of Linda Creek (sections 6.10). The position and geometry of the sediments in these environments suggests that the nature and location of deposition were topographically controlled. Thick sequences usually occur where lakes formed between the ice margin and hillslopes, and in small ice-dammed valleys. Although similar environments can occur in supraglacial positions, they are not as long lived, and are not as likely to accumulate thick sequences of sediment.

The dominant and diagnostic sediments of subaqueous ice-contact environments are lacustrine. They accumulated in small pools and lakes between glacier margins and tributary valleys. Because they formed in ice-dammed lakes in close proximity to high sediment fluxes, the sediments are highly variable and consist of low energy (eg laminated silt) to high energy (eg mass flows), depending on the supply and availability of meltwater and debris.

7.4.2 Lacustrine environments.

Two types of lacustrine sediments that can be distinguished are ice-contact and/or ice-dammed, and moraine dammed. The ice-contact lacustrine environment is relatively ephemeral and consists of laminated silts interbedded with subaqueous sediment flows, ice-rafted diamictons and outwash gravels. Moraine-dammed lakes in the King Valley have formed long sedimentary records (>10^2 yrs.) in deep (> 50 m) basins producing uninterrupted sequences of laminated silts and muds.

Two major sequences of moraine-dammed lake sediments are the Nelson Formation which accumulated behind the King Formation terminal moraine, and the lake sediments that underlie the Chamouni Formation. Both are believed to represent non-glacial events in the valley, although ice may have been present in the region when they accumulated. The position of accumulation of these sediments was controlled by topography and the terminal position of ice advances. It appears that when a terminal position of an advance coincided with a narrow part of the valley and sufficient deposits accumulated, a dam was formed. They probably accumulated in a position similar to lakes Pukaki, Ohau and Tekapo, which have formed behind terminal moraines of the Otira Glaciation in New Zealand (Pickrill and Irwin 1983). These lakes are presently fed by meltwater streams from a remote ice source.

7.4.3 Glaciofluvial sediments.

Two distinct facies of outwash gravel in the King Valley are coarse, massive, clast-supported gravel without much matrix, and horizontal and cross-bedded sandy gravel.

Massive gravels are more common and constitute the underlie outwash plains in the middle and lower King Valley. They are similar to those described as "Scott type" facies by Miall (1978). Scott type outwash facies consist of >90% gravel that is dominated by massive gravel almost to the exclusion of cross-bedded gravel (Rust 1978). They are representative of the proximal reaches of braided outwash streams (Miall 1978).

Bedded sandy gravels and sands are relatively uncommon and are restricted in area to the lower King Valley where they form the distal outwash gravels of the Blackwood Formation. All other outwash deposits in the King Valley are dominated by Scott type facies irrespective of their position of deposition relative to the ice front. This is because the valley is narrow and there was insufficient space for inactive areas to develop in the outwash streams. Consequently, outwash streams had high competence throughout their reaches and finer sediments were removed through the King Gorge in the West Coast Range.

7.5 Lithological stratification.

A pattern of increasing erratic content with increasing depth was observed in several sections through the glacial deposits in the King Valley. Initially it was thought that the lithological stratification could be explained entirely by dilution of the erratic content of surface sediments by locally derived, non erratic rocks. However, occurrence of the same pattern in glacial sediments remote from non-erratic rock sources suggested that the stratification was inherited from the processes of transport or deposition of debris by the King Glacier. The results of pebble counts from two sections in which lithological stratification was observed are summarised in figures 6.5 and 6.58.

Figure 6.5 illustrates a series of five stone counts in a section of early Pleistocene glacial sediments that records supraglacial deposition on melting glacier ice. The geometry, pebble fabric, and lenses of stratified sand suggest that the basal sediments are melt-out tills overlain by thin multiple sediment flows. The sediment flows are overlain by delta foreset sands, a thin sediment flow and laminated silts that were eroded and buried by post glacial slope deposits derived from adjacent slopes. The lithology of the melt-out till is mixed and has a high content of erratic Jurassic dolerite and Permian rocks. The proportion of erratic clasts decreases suddenly in the sediment flows which consist mainly of quartz, Ordovician conglomerate and Cambrian volcanic pebbles. The overlying slope deposits contain no erratic pebbles and are entirely derived from adjacent slopes which consist of Devonian quartzite and Ordovician conglomerate.

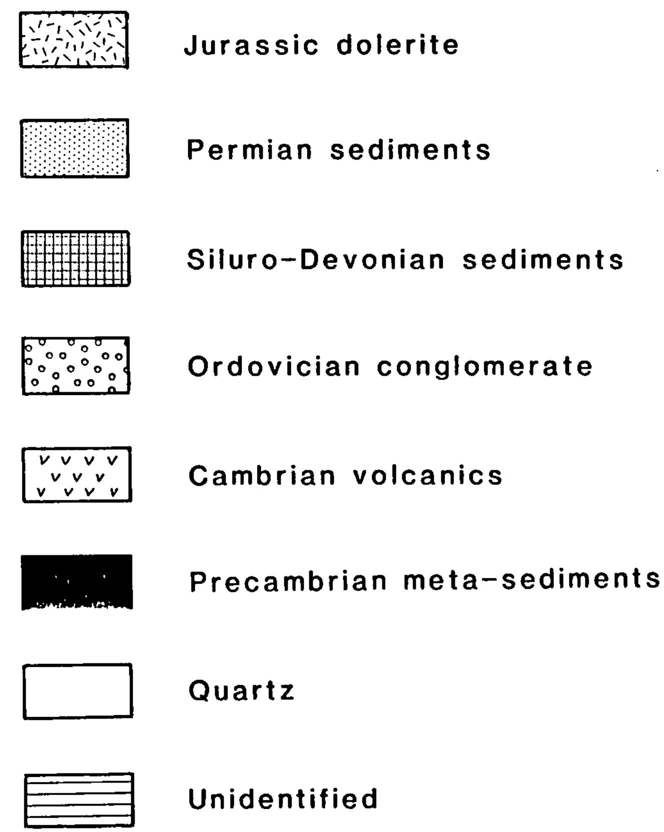
Figure 6.58 shows a section through the Blackwood Formation end moraine that rests on 43 m of interstadial laminated lake sediments. Four stone counts show that the sediments consist of a mixture of locally derived, West Coast Range and Eldon Range lithologies. There is a gradual upward decrease in erratic content to the top metre of the section where the proportion decreases sharply.

The lithological stratification may have two causes. Firstly, it could have been due to dilution of erratic clasts in near surface glacigenic sediments by local extra-glacial sediment fluxes. Secondly, the stratification could be inherited from rock from different sources becoming entrained and transported at different levels in the glacier.

The amount of a given rock type in till depends on several factors that include the thermal conditions of the glacier sole (Boulton 1972a, 1972b, 1974), the area of outcrop, the durability of the rock to processes of erosion (Clague 1975), distance of transport, and progressive deposition and dilution (Peltoniemi 1985). Studies of till composition have suggested that an important means of producing vertical variations in lithology is by the englacial stratification of debris eroded from different subglacial outcrops (Boulton 1970a, 1970b, Rose 1974, Ehlers 1981, Rappol and Stoltenberg 1985). Such a mechanism would produce a decrease in erratic content with depth, and does not explain the opposite pattern observed in the King Valley.

The explanation presented here suggests that the King Valley pattern is due to a lithological contrast between supraglacial debris derived from the valley sides and debris carried within and beneath the glacier. This difference is considered to be due to differences in outcrop lithology above and below the equilibrium line. Figures 7.2 and 7.3 summarise the model. The flow paths are based on Boulton's (1978) model of transport paths of debris through cirque and valley glaciers.

Lithological contrasts between supraglacial and englacial debris occur because of the position of the equilibrium line altitude (ELA) of the King Glacier relative to the altitude and position of the erratic rock types of the Eldon Range (Fig. 7.3). Colhoun (1985a) calculated the ELA for the King Glacier during the last or Margaret Glaciation to be at 835 m. During the earlier, more extensive glacial advances the ELA would have been considerably lower. The altitude of the unconformity at the base of the Permian sediments on the Eldon Range is 1030 m. (Read 1963). Therefore the sources of Permian sediments and Jurassic dolerite were always above the ELA's of glaciers in the King Valley (Fig 7.3). For this reason these erratics would have been transported in an englacial and subglacial positions regardless of whether they were eroded in a


Jurassic dolerite
Permian sediments
Siluro-Devonian sediments
Ordovician conglomerate
Cambrian volcanics
Precambrian meta-sediments
Quartz
Unidentified

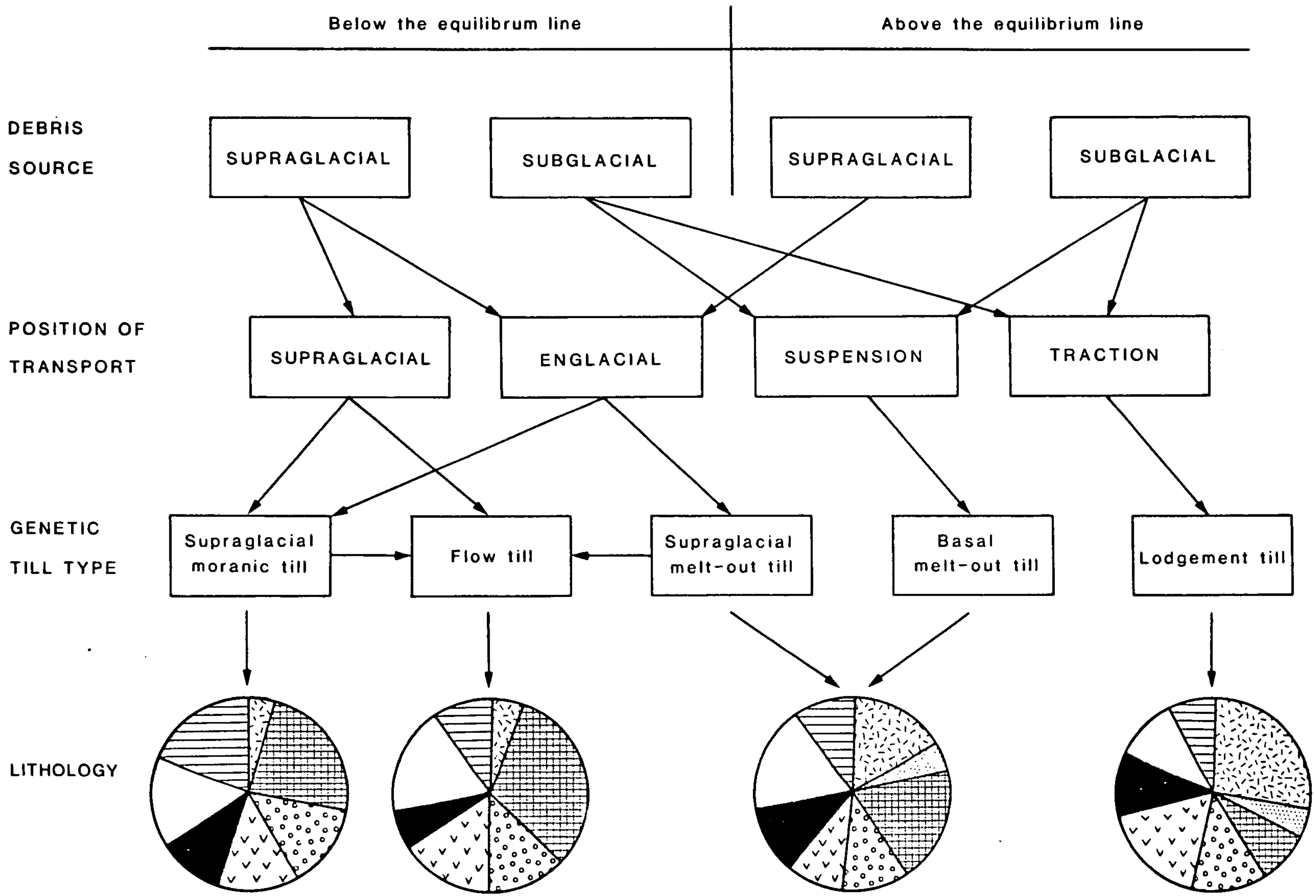


Fig. 7.2. Model of lithological stratification of glacial sediments in the King Valley. Modified from Boulton (1978).

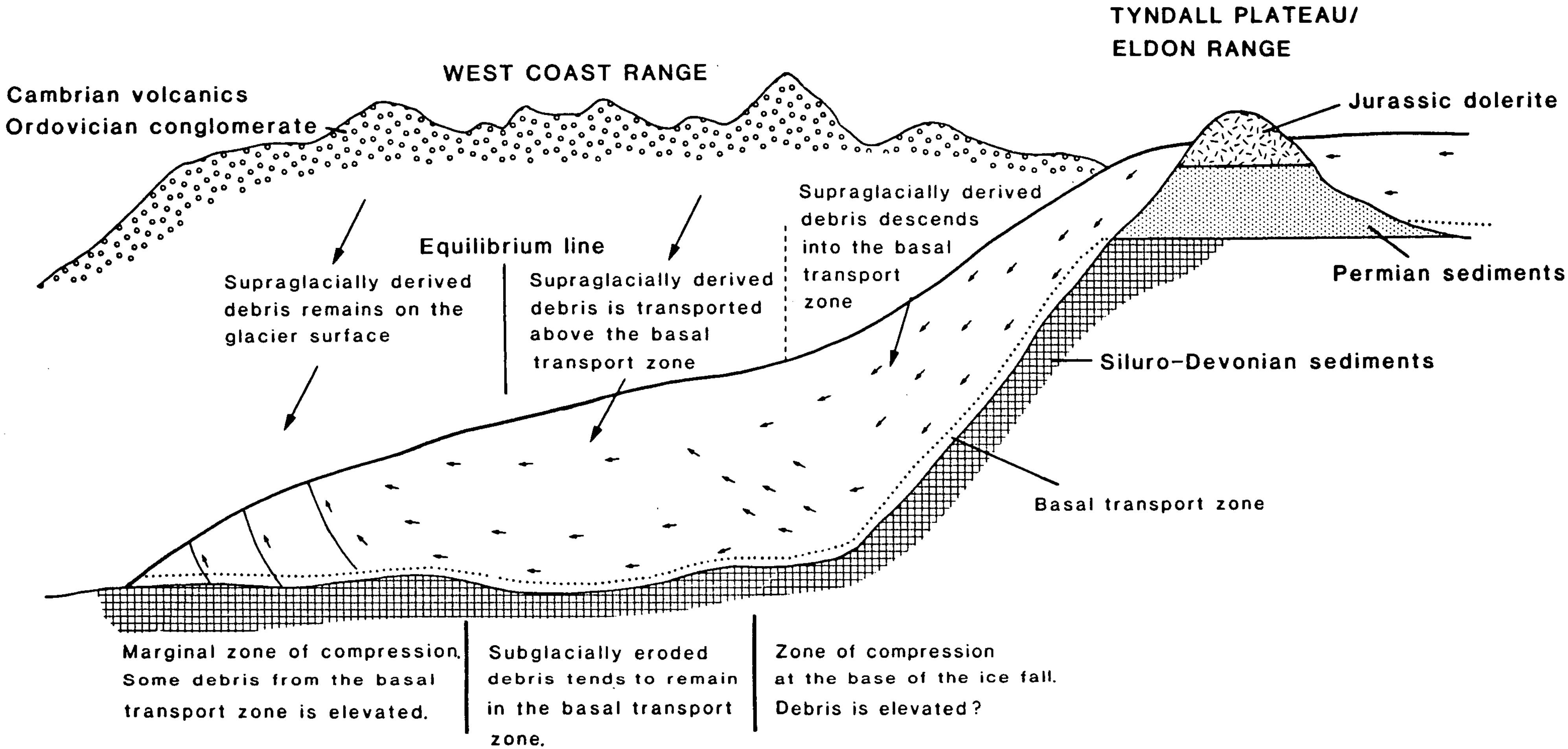


Fig. 7.3. Location of erratic and other rocks in the King Valley relative to the position of the Equilibrium Line Altitude of the King Glacier during the last or Margaret Glaciation. Symbols for rock types are the same as those used in Fig. 7.2.

subglacial position or accumulated on the glacier surface. Other debris that accumulated on the glacier surface below the equilibrium line was carried in a supraglacial position. This debris, would therefore have been deposited above that transported in englacial and subglacial positions. The lithology of the major rock types that would have contributed to supraglacial debris below the equilibrium line were Ordovician conglomerate and Cambrian volcanics from the West Coast Range. Consequently, these rock types dominate the composition of some surface till deposits.

Thus, ice terminal and stagnant ice depositional environments in the King Valley favour deposition of a supraglacial till facies dominated by siliceous, non erratic lithologies.

If the depositional system of the King Glacier was functioning as the model describes, there would be an absence of erratic clasts from all supraglacial sediments. Observations of the sediments suggests this is not the case (Chapter 6). Three possible reasons for this inconsistency are; mixing of debris within the glacier, uneven or absence of supraglacial debris cover on parts of the glacier surface and mixing of the debris during deposition.

Although glaciers are often assumed to be stratified transportation mediums with little or negligible diffusion of debris (eg. Clague 1975), subglacial observations suggest it is possible for material from the basal transport zone of glaciers to elevated to higher levels (Boulton 1978). Although there is evidence that suggests that basal debris was elevated within the King Glacier (section 7.6) the elevated debris could not have become mixed with supraglacial debris until it was released onto the glacier surface (Fig. 7.2).

Many sequences of deposits in the King Valley do not have a supraglacial till facies. This is probably because parts of the King Glacier lacked substantial supraglacial debris which tends to be concentrated on ice streams that pass close to rock walls. In such deposits lithological stratification would be absent or limited to a very thin bed. This may explain why the pattern is not observed in some sections of the valley but tends to occur in sections through lateral and latero-frontal moraines where supraglacial debris is normally concentrated.

Recent studies of ice terminal and supraglacial depositional environments have stressed the importance of secondary processes that rework debris initially deposited as till (Lawson 1979; Eyles 1979). These secondary processes cause mixing of supraglacial and englacial debris that accumulate on melting ice and subglacial surfaces. The occurrence of sediment flows in many supraglacial sediments in the King Valley (Fig 4.5 and 4.58) suggests that mixing of this type was a common process in the ice terminal and deglacial environments. Thus these secondary processes may have caused limited mixing of the siliceous supraglacial debris with the erratic rich englacial debris.

Given that the discharge of supraglacial debris in ice sheets is negligible (Boulton 1978), lithological stratification by the mechanism described here is a phenomenon mainly of alpine glaciation. Presumably because most detailed studies of lithology of glacial sediments have been confined to deposits of the continental glaciers that covered North America and Europe, the lithological stratification related to supraglacial debris has not been described before.

However, of the few studies of the composition of tills of alpine valley glaciers, none record stratification of lithology. Clague's (1975) study of till deposited by glaciers of the Rocky Mountains found that they had a vertically homogeneous composition. Clague suggests that the lithologies were mixing by englacial meltwater and to a lesser extent by active ice transport (ibid p. 730). Contrary to Clague's study, this study suggests that stratified ice transport was the major control of till composition.

The patterns of lithology in sediments in the King Valley demonstrate that given a knowledge of the geographic distribution of rock types, it is possible to reconstruct the flow paths of debris in the King Glacier and that specific till types may have a characteristic lithological composition.

7.6 Discussion

Several questions arise from the consideration of the patterns and processes of sediment genesis inferred from the deposits of the King Valley. The two most important are the origin of the widespread supraglacial sediments and the reason for the lack of lodgement tills. These questions appear to be related because both may involve the elevation of subglacial debris to higher levels within the glacier. To determine the significance of the two observations elements of the debris transport paths and dynamics of the King Glacier are reconstructed. This reconstruction suggests that most of the supraglacial sediments are subglacially derived and form a sediment association that is generally considered characteristic of glaciers that have a sub-polar thermal regime. The significance of this suggestion is explored with reference to the inferred thermal conditions and flow dynamics of the King Glacier. Because it is likely that the King Glacier had a temperate thermal regime, it seems necessary to invoke other mechanisms to explain the possible elevation of basal debris into higher transport levels.

Supraglacial sediments can have two origins. One includes debris derived from the valley sides and deposited on the glacier surface by rockfalls, alluvial fans and avalanches. Such debris is transported passively, either in a englacial or supraglacial position depending on whether it accumulated above or below the equilibrium line. Debris that follows this transport path does not undergo a phase of traction and has distinct material properties. It is characterised by a predominance of angular oxidised clasts, a unimodal particle size, poor sorting, a mud content below 15%, normal consolidation, high porosity and a low bulk density (Eyles 1979). This material has been called supraglacial morainic till by Boulton and Eyles (1979).

Another source of supraglacial debris is basal debris that is transported to higher levels and accumulates on the glacier surface as it melts. Because this material undergoes a phases of traction the material properties are very different from supraglacial morainic till and are similar to subglacial sediments (Boulton 1978). Debris from this source forms what has been called the supraglacial landform and sediment association (Table 7.1), and is characteristic of glaciers with a complex, subpolar thermal regime (Boulton and Paul 1976; Boulton 1979).

Because the glacial system of the King Valley was one of ice cap and outlet glacier, there were few sources from which supraglacial debris could fall onto the glacier surface. The main source

Table 7.1 Classification of glacial landforms and sediments.

<table>
<tr><th>Landsystem</th><th>Sediment association</th><th>Source of glacial debris</th><th>Principal till types</th></tr>
<tr><td>1. Glaciated valley landsystem May be superimposed on 2 and 3</td><td rowspan="2">(a) Supraglacial sediment association</td><td>Supraglacial - nunataks and valley sides</td><td rowspan="2">Supraglacial morainic till
Flow till
Melt-out till</td></tr>
<tr><td>2. Supraglacial</td><td rowspan="2">Subglacial</td></tr>
<tr><td>3. Subglacial proglacial landsystem</td><td>(b) Subglacial proglacial sediment association</td><td>Melt-out till
Lodgement till
Lee side till</td></tr>
</table>

after Boulton and Paul (1976)

was the steep rock walls of the West Coast Range. Debris from this source would be concentrated in narrow streams that would have covered little of the glacier surface. Sediments from this origin are limited to small areas in the King Valley and are identifiable by their lack of erratic rock types (see section 7.5). This together with the abundance of supraglacial sediments in the King Valley that do not have material properties anything like supraglacial morainic till strongly suggests that the supraglacial sediments have a subglacial origin and have undergone a traction phase of transport.

The problem therefore, is to determine how the King Glacier became charged with englacial debris from the basal transport zone.

Studies of the genesis of sediments at the margins of glaciers agree that ice thermal conditions are a fundamental control of the mechanisms and position of glacial deposition (Boulton 1972a, 1975, Paterson 1981, Shaw 1977). The elevation of basal debris into englacial positions has been claimed to be associated with glaciers that have a complex thermal regime (Boulton 1968, 1972b, 1978). Lodgement till on the other hand is considered to be the predominant mode of deposition of temperate glaciers (Boulton 1972b). Resolving the significance of the patterns observed in the King Valley therefore depends on an understanding of how thermal regime controls the position of transport and deposition of debris. It is therefore necessary to know how thermal regime controls the nature and dispersal of sediments, what characteristic sediments are related to the different thermal regimes and, if it is possible to reconstruct the thermal regime of an ancient glacier from palaeogeographic, climatologic or sedimentologic data.

Paterson (1981) identified two fundamentally different types of glacier, temperate and polar. Temperate glaciers are at pressure melting point throughout except where a seasonal cold wave of precipitation is added. Polar glaciers have three different thermal conditions; all ice below melting point, a layer of cold ice underlain by ice at melting point, and all ice below melting point except at the glacier base where ice reaches melting point (ibid p. 192). Boulton (1972a) identified three types of glaciers. Polar glaciers in which ice temperature is always below melting point, circumpolar glaciers in which ice reaches melting point for one or two months of the year, and temperate glaciers in which ice is rarely below melting point. Andrews (1975) classified glaciers in a similar manner but described circumpolar glaciers as sub-polar.

Observations from temperate glaciers reveal that basally derived debris is transported almost exclusively within the lowermost metre and deposited as lodgement till. In cold glaciers the debris is transported at higher levels and is deposited as supraglacial till (Kamb and La Chapelle 1964; Boulton 1970a, 1972a). These empirically known patterns were put into a theoretical context by Boulton (1972b). He suggested that glacier thermal regime controls the location and nature of sedimentation by determining the thermal boundary conditions that define the sedimentary processes. Boulton defined three thermal boundary conditions, net basal melting, a balance between melting and freezing, and net meltwater freezing onto the glacier sole (ibid p. 2). Deposition during the first two conditions occurs subglacially, producing mainly lodgement till. During the third condition debris is transported to an elevated position and is deposited as supraglacial melt-out and flow tills (ibid p. 13). According to this theory, temperate glaciers deposit most debris subglacially as lodgement tills, and polar and subpolar glaciers deposit most debris supraglacially as melt-out and flow tills. Debris is transported into the higher, englacial, transport zone in glaciers with a complex thermal regime because alternating zones of basal melting and freezing of the glacier sole to the bed creates zones of compression (Nye 1952). The compression generates an upward component in flow that facilitates the upward transport of debris.

Did the King Glacier have a complex thermal regime similar to modern subpolar glaciers? The thermal regime of glaciers is determined by several environmental variables that include mass balance, ice thickness, ice velocity and climate. It is not possible to reconstruct the glaciological variables of the King Valley. However, its location in a temperate maritime climate at 42°S suggests that the climate in western Tasmania is likely to have been dominated by heavy winter snowfall and intense summer melt. Such climates are likely to support temperate glaciers (Paterson 1981). The King Glacier was probably temperate throughout the Quaternary and since it would not have been frozen to its bed it is unlikely that the thermal regime of the ice caused the elevation of basal debris.

The basis of the hypothesis of thermal regime controlling the position of transport and deposition of debris is that alternating zones of freezing and melting cause flow extension and compression that drives an upward flow component. Thermal regime is but one means of generating flow compression. Several alternative means include lateral compression and longitudinal compression, both of which depend on the mechanics of flowing ice, valley form or bed topography. Although the freezing-on of large amounts of regelation ice may also elevate basal debris, it is extremely unlikely to have occurred in sufficient quantities to have elevated much debris into an englacial position (G. S. Boulton pers. comm. 1988).

Transverse compression usually occurs where two ice streams flow together. Although ice streams almost certainly did coalesce in the upper parts of the King Glacier, the absence of medial moraine forms suggests that transverse compression was not significant and did not elevate much basal debris. The zones were probably too localised to elevate debris in the amounts that would be needed to produce the extensive supraglacial sediments in the valley.

Longitudinal flow compression occurs in a variety of ways including when the rate of flow becomes slower in the terminal area, where the glacier flows against bedrock or against old moraines. Flow compression is common in the outer areas of temperate glaciers. However, the compressed zone is usually quite narrow and basal debris is seldom lifted to the glacier surface. If it is, the area of glacier ice covered by supraglacial debris is small. This mechanism is unlikely to have resulted in the widespread occurrence supraglacial sediments in the King Valley.

Basal debris is known to become elevated during glacier surges (Clapperton 1975, Sharp 1985). Clayton and Moran (1974) have described similarities between supraglacial landforms in Dakota and those currently forming on modern surging glaciers in Alaska. During surges the terminal zone experiences intense longitudinal compression. Sharp describes multiple englacial debris bands in a surging Icelandic glacier that he attributes to shearing of thrust-faulted basal debris-rich ice. The englacial debris bands have a lateral continuity of over 1 km and result in a belt of hummocky and kettled topography, and a supraglacial sediment association forms during quiescent phases (Sharp 1985). However, to what extent the elevation of basal debris in the King Glacier was related to surge behaviour remains unknown though the bed topography may have been conducive to thickening, instability and surge behaviour.

In the King Valley major zones of longitudinal compression are likely to have occurred where ice flowed from the Tyndall Plateau onto the valley floor. The ice fall from the Tyndall Plateau and highland area behind the Eldon Range (Fig. 7.3) to the bottom of the King Valley is a 600 m decrease in altitude in about 1.5 km. Because most of the King Glacier passed over these falls, it is likely that a zone of intense flow compression developed at the base of the glacier near the foot of the ice fall. Such zones of compression are associated with an upward flow vector and would transport debris into higher positions in the glacier. The precise mechanism of debris elevation in this situation is unclear. However, it is likely to involve shear planes or folding of ice in the basal transport zone. The shear plane hypothesis was formulated by Nye (1952). He suggested that in zones of compressive flow "slip lines" that define the direction of maximum shear favour the upward movement of ice and sediment (ibid p. 88). These zones of shear are tangential to the glacier bed. Their existence is confirmed by the observation of thrust faults and debris bands on the surface of glaciers in zones of compressive flow (Nye 1952; Gunn 1964). However, Weertman (1961) and Hooke (1968) have disputed the ability of shear zones to transport subglacial debris into englacial positions.

Folding of the basal debris zone is common in areas of flow compression of temperate glaciers. Such folding results in the repetition of basal debris layers and thickening of the basal debris zone (N. Eyles pers. comm. 1988). Combined with an upward flow vector, folding may result in the elevation of parts of the basal debris zone into an englacial position so that when the glacier melts subglacially derived debris is released on the glacier surface.

The development of a supraglacial sediment association may well be associated with subglacial relief. This relationship is not unknown, as supraglacial landforms and sediment associations have a consistent relationship to other landforms including bedrock obstructions and escarpments (Paul 1983).

Although the actual mechanism of elevation of basal debris in the King Glacier cannot be clearly established, the circumstantial evidence and the general lack of lodgement till suggests the subglacial topography of the King Glacier was the most likely cause. If this interpretation is correct, the similarity of the supraglacial sediment association to that associated with complex, subpolar glaciers brings into question the reliability of determining the thermal regime of a glacier from diagnostic sediment associations (Boulton 1972a, 1978, Eyles and Slatt 1977).

SUMMARY

- The dominant processes inferred from deposits in the King Valley involve meltwater flow and reworking of primary tills by sediment flow and other processes of mass movement. Primary tills deposited from active and melting ice (lodgement and melt-out till) are relatively rare.

- The major ice-contact sedimentary environments are supraglacial and ice marginal. The supraglacial sediments are mainly derived from the basal debris zone and form a sediment association that is regarded as characteristic of with subpolar glaciers.

- Sediments derived from the valley sides (supraglacial morainic till) have a distinctly local lithological composition. They generally occur above the subglacially derived supraglacial till which has a higher erratic content.

- While the lack of lodgement till and the importance of the subglacially derived supraglacial sediment association has been interpreted as suggesting a subpolar thermal regime, this interpretation does not appear to be valid for the King Valley. Reconstruction of the flow paths of the King Glacier brings into question the assumption that landform and sediment associations are solely determined by the thermal regime of glaciers.

CHAPTER 8.

CHRONOLOGY, ENVIROMENTAL CHANGE AND CORRELATION.

8.1 Introduction.

This chapter outlines the chronology and estimates the ages of the glacial events in the King Valley. It discusses the environmental history of the King Valley and western Tasmania, and compares the chronology with other glacial sequences in Tasmania. Comments are made on the prospects of correlating the glacial events with those recorded from Southern Hemisphere middle latitude areas.

8.2 Chronology of the glacial deposits of the King Valley

The tentative ages of the glacial formations recognised are summarised by Figure 8.1. This chronology is based on estimates of ages using several methods. Most of the estimates are minimum ages and should be regarded as a first approximations.

The maximum age of the Long Marsh Formation has been determined by a radiocarbon date of 12,250 ± 90 yrs. B.P. (SUA 2415) from a large *Eucalyptus* branch beneath alluvial gravel (Fig. 4.76). The date suggests that the upper part of the King Valley supported a *Eucalyptus* forest with relatively large trees by this time. From this it is inferred that the valley was largely ice free.

The age of the Dante Formation has been determined by ^{14}C dating that suggests it post dates 18,800 yrs. BP. Although this date has been given the status of "the" date of the maximum of the Last or Margaret Glaciation, the section is not without its problems. The existence of unweathered, Last Glaciation tills (the Chamouni Formation, >48,000 yrs. B.P.) outside the limits of glacigenic sediments of the Dante Formation demonstrated that the Dante Formation does

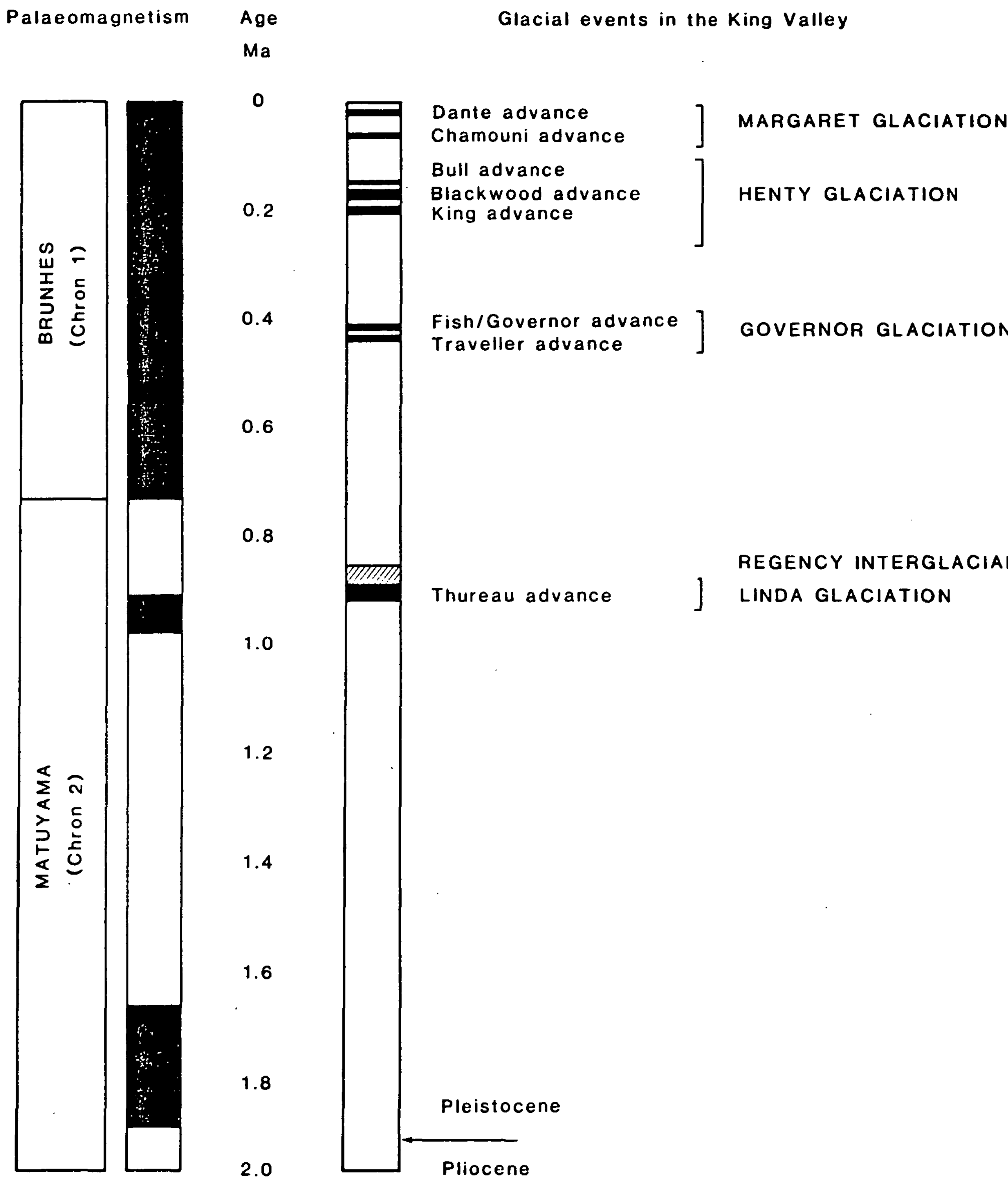


Fig. 8.1. Tentative ages of the glacial advances in the King Valley.

not necessarily record the maximum extent of ice during the Last Glaciation. There are three possible interpretations of the stratigraphy:

1. the Chamouni Formation is slightly older than the Dante Formation;
2. the Chamouni Formation represents an early Last Glaciation ice advance;
3. the Chamouni Formation was deposited prior to the Last Glaciation.

All these alternatives suggest that the basal sediments of the Dante section (Fig. 4.75) are not of the Comstock Glaciation as suggested by Kiernan (1983a) and Gibson *et al.* (1987) but are outwash gravels of the Chamouni Formation. The soil developed on these gravels has no unconformity and records an alpine humus soil rather a lowland brown or podsol interglacial type soil that would have developed on Comstock gravels. The most logical interpretation is that the soil developed on the Chamouni outwash gravel during a period when the area was ice free but was sufficiently cool to allow the development of only an alpine flora. While this interpretation does not assist with the assessment of the age of the Chamouni Formation, the development of a soil records the period that separates the Dante and Chamouni Formations (the King Interval of Gibson *et al.* 1987). A dates from a fossil *Donatia novae-zelandiae* that grew on the palaeosol at Dante Rivulet (Fig. 4.75) suggest that the basal gravels are older than 21,180 yrs. BP, which can be regarded as a minimum age for the Chamouni Formation. Further down the valley a date of $48{,}700^{+2900}_{-2100}$ yrs. BP (SUA 2599) was obtained on drifted wood that apparently formed part of the Chamouni Formation outwash gravel. The date suggests the age of the formation is beyond ^{14}C dating and therefore more likely to have been deposited by an ice advance that occurred substantially before the Dante advance. The Chamouni Formation is interpreted as younger than the Henty age formations of the King Valley because its deposits lie geographically closer to the sources of ice and the deposits are considerably less altered by weathering. At present the evidence for the Chamouni advance is equivocal. The two possible interpretations are that it represents an early Last Glaciation ice advance or an ice advance that occurred between the Margaret and Henty Glaciations. Future research may resolve this problem.

The age estimate of the Bull Formation is dependent on its close geographic relationship, and inferred temporal relationship, to the Blackwood Formation. It is therefore considered to be of Henty Glacial age though there is little exposure from which to deduce the characteristics of the deposits which may prove to be considerably younger.

Although there is no reliable absolute date for the Blackwood Formation, in the lower King Valley the two ^{14}C dates of 32,800 $^{+400}_{-700}$ yrs. B.P. (SUA 2392) and 39,300 $^{+800}_{-700}$ yrs. B.P. (SUA 2393) that were probably contaminated by modern carbon suggest it is beyond the range of radiocarbon dating (section 3.3). The age of the sediments can be assessed by comparing the weathering rinds with those in the dated Dante Formation. The mean weathering rind thickness for the Blackwood Formation is 15.6 mm, which is over 8 times that of the Dante Formation (1.9 mm). Assuming a linear weathering rate and given the age of the Dante Formation (18,800 yrs.) the weathering rind thicknesses suggest that the Blackwood Formation is at least 150,000 years old.

The age of the King Formation can be estimated by assessing the period of time represented by the Nelson Formation that lies between the King and Blackwood formations. The period of time represented by the 43 m of Nelson Formation lake sediments of the have been estimated as representing about 4300 years assuming similar sedimentation rates to Holocene proglacial lakes in New Zealand (section 6.9). Because the top of the lake sediments are eroded this estimate is probably conservative. Taking the age of the Blackwood Formation as 150 ka, the age of the King Formation is approximately 154,000 yrs. Although this estimate cannot be regarded as precise, it suggests that the deposits are not younger than the last interglacial in age and are not extremely old.

The age of the Governor Formation can be estimated by comparing the thickness of weathering rinds with those of the King Formation. This direct comparison is only possible because the contact between the King and Governor formations can be observed in two places. At these sites the King Formation has mean weathering rinds of 5.1 and 5.3 mm and the Governor Formation has ones of 14.3 and 17.3 mm. The difference suggests that the Governor Formation is 2.7 to 3.4 times older than the King Formation. This gives an estimated age of between 415,000 and 520,000 years old. Judged by the oxygen isotope record (Shackleton and Opdyke 1973) the magnitude of the estimated unconformity between the King and Governor formations suggests they were deposited during different glaciations.

The Fish Formation has been shown to differ little in stratigraphic age from the Governor Formation and therefore is regarded as being of approximately the same age.

The Traveller formation is separated from the Fish and Governor formations by the Baxter Interstadial deposit. The period of time represented by the Baxter Interstadial deposit is unknown. The time gap between the deposition of the Traveller and Fish formations cannot be estimated by weathering because the sediments are largely siliceous and contain no dolerite. The Baxter Interstadial sediments have a normal detrital remanent magnetisation which indicates that it belongs in the Brunhes Chron which is in accordance with age estimates for the Governor Formation.

The age of the Thureau Formation is more difficult to estimate. It is known to have a reversed detrital remanent magnetisation and therefore is probably older than 730,000 yrs. However, because the sediments at the type section of the Thureau Formation appear to cross a palaeomagnetic boundary (M. Pollington pers. comm. 1987) it may be that they contain part of the Jaramillo event in the Brunhes Chron, which occurs at 910,000 yrs. B.P. (Bowen 1978).

8.3 Environmental change, evidence from the King Valley.

One of the major problems in discussing environmental changes during the Quaternary in Tasmania, as elsewhere, is the limited evidence, as very little of the Quaternary is represented by sediments (Fig. 8.1). From the available evidence it seems that environmental conditions varied little in preglacial times. Once glacial climatic conditions began to influence the region the environment experienced high sediment fluxes associated with multiple ice advances. There were also dramatic changes in flora between glacial and interglacial stages. Because the tectonic environment of western Tasmania was relatively stable during the Quaternary, most environmental change can be directly attributed to climatic change rather than increase in altitude of the mountains.

Glaciation is used in both a climatic and in a stratigraphic sense. In the climatic sense, in the King Valley a glaciation is an event of cooling that leads to the formation of glaciers. In a stratigraphic sense, it is necessary to distinguish the event from the deposits that formed as a consequence of the event (Suggate 1965b). In this thesis a glaciation is defined as a number of glacial advances that occur in close temporal proximity to each other.

Interglacials are non glacial periods in which either the length or degree of climatic amelioration permit the development of forest indicative of climate at least as warm as was attained during the post-glacial climatic optimum (West 1961). In the King Valley interglacials are recognised as biostratigraphic units defined by the presence of pollen or plant macrofossils that record the development of temperate rainforest.

Stadials are cooler periods within glaciations during which glaciers are more extensive and are separated by interstadials which are periods of climatic amelioration within glacial events (Bowen 1978). Interstadial climates are generally considered to have been cooler than the Holocene optimum. During interstadials the temperature probably rose to about halfway between the minimum temperature of the Last Glaciation and the maximum warmth of the post glacial climatic optimum (West 1961, Suggate 1965b). In the King Valley stadials are recognised by glacier advances and defined as morphostratigraphic and/or lithostratigraphic formations. Interstadials are recognised as biostratigraphic and/or lithostratigraphic units on the basis of having heath and shrub vegetation assemblages and/or accumulations of non-glacial sediments.

8.3.1 The Pre-glacial Environment.

Recognition of the preglacial environment is based on the absence of evidence for glaciation. It can be considered a stable period of subtropical-temperate climate with a slow cooling trend from the middle to late Tertiary. One of the difficulties in discussing the pre-glacial environment is that the antiquity of glaciation in Tasmania is not known, it may be early Pleistocene or Late Tertiary. The tillites of the oldest glaciation at Lemonthyme Creek are overlain by sediments that contain a palynoflora of Tertiary age that indicates temperate rainforest conditions.

The geological environment of the Tertiary was dominated by terrestrial sedimentation in a number of block faulted grabens, basalt flows and marine sedimentation (Gill 1962). Tertiary basalt flows are absent from the central West Coast Range but the nearby Macquarie Harbour was a site of sedimentation in which over 230 m of clays, sands, silts, lignites and conglomerates accumulated and are believed to vary in age from early Eocene to Late Pliocene-Early Pleistocene age.

The only preglacial sediments in the King Valley are the fluvial sediments of the Gormanston Formation which are preserved as an inlier in Thureau Formation glacigenic sediments in the Linda Valley.

Tertiary pollen and macrofossils of tropical and subtropical plant taxa known in Miocene and Oligocene sediments (Gill 1962) show evidence for higher temperatures than present. A cooling trend appears to occur in the Late Tertiary (Hill and Macphail 1983). The plant fossil composition at Pioneer in northeastern Tasmania describes a complex temperate rainforest dominated by *Nothofagus*. From a count of 618 pollen grains Hill and Macphail (1983) found that 47% were *N. emarcidus/heterus*, an extinct species of the *brassi* group. The Gormanston palaeosol also contains forest taxa that are now extinct including 13% *N. brassi* type pollen (Kiernan 1980). At Regatta Point, about 20 km west of the King Valley, an eroded block of organic silt that is probably no older than Pliocene in age, contains pollen of temperate rainforest taxa some of which are now extinct (Hill and Macphail 1985).

Traditionally in Tasmania the absence of extinct temperate rainforest species is often referred to as the difference between the Tertiary and the Pleistocene and many floras such as the Gormanston one are claimed to have "Tertiary affinity" (eg Kiernan 1980 and Colhoun 1985a). The reality is that the Tertiary, and particularly the later part of the Tertiary, is characterised by a lack of knowledge about the composition or temporal range of the flora. At present it is impossible to estimate the age of deposits from floristic characteristics. The range of many of the species that are often claimed to have "Tertiary affinity" is not known. The identification of *Quintinia psilatispora* and *Gothanipollis perplexus* in Pleistocene sediments in the King Valley (section 4.4) has considerably extended their temporal range (G. van de Geer and E. A. Colhoun pers. comm. 1987).

Little is known about the late Tertiary preglacial vegetation other than that many species became extinct during the late Tertiary and early Pleistocene. To what extent the extinction of numerous forest taxa recorded in late Tertiary sediments is related to cooling temperatures and/or glaciation remains unknown.

8.3.2 The Lemonthyme Glaciation.

Evidence for the Lemonthyme Glaciation consists of drill cores through sediments at Lemonthyme Creek in northern Tasmania. The cores show sequences of laminated muds overlying and interbedded with tillite (Paterson 1965). The silts contain pollen that indicates cool temperate rainforest dominated by *Nothofagus* existed after the deposition of the tillite (Paterson *et al.* 1967). Over half of the *Nothofagus* pollen grains are species that are now extinct, suggesting that the flora has closer affinity to those described from Pioneer (Hill and Macphail 1983) and Gormanston (Kiernan 1980) than to any Quaternary interglacial flora.

Because the Lemonthyme Glaciation is known only from subsurface information at Lemonthyme Creek, neither the extent of ice, nor the environment at the time of deposition are known. No glacial deposits of this antiquity are known in the King Valley or elsewhere in Tasmania. Their age and full significance remains to be evaluated.

8.3.3 The Linda Glaciation.

Deposits of the Linda Glaciation are clearly the most extensive glacial deposits in western Tasmania. This does not agree with oxygen isotope records from the southern ocean, most of which suggests that maximum ice volumes were attained at about stage 6 (c140,000 yrs. BP, Shackleton and Opdyke 1973). The Linda deposits consist of a wide range of sediments dominated by flow tills, laminated lake sediments and ice-rafted tills. The lack of melt-out till and abundance of outwash gravel suggests that the glaciers were temperate and the depositional environments characterised by large fluxes of meltwater. The altitude attained by erratics on the hillsides suggests that the King Glacier was at least 400 m thick in the middle part of the valley near the Thureau Hills.

Although there is some suggestion of multiple ice advances in the Thureau Formation, evidence to substantiate this and suggestions that deposits of the Linda Glaciation contain deposits of several glaciations (Kiernan 1983b; Gibson *et al.* 1987) has not been forthcoming.

No glacial floras of Linda Glaciation age are known in Tasmania. Deposits of the Regency Interglacial conformably overlie the Thureau Formation and suggest that the environment immediately following deglaciation was dominated by aquatic and subalpine taxa. The Regency Interglacial deposits are the only ones known to occur between the Linda Glaciation and subsequent middle Pleistocene ice advances.

8.3.4 The Regency Interglacial.

The Regency Interglacial consists of 0.9 m of drifted wood, leaves and organic silty sand that conformably overlie Thureau Formation outwash gravels. Pollen analysis of these sediments indicates the development of temperate rainforest dominated by *Nothofagus cunninghamii, Phyllocladus aspleniifolius* and *Lagarostrobos franklinii*. (Fig 4.1). The abundance of tree ferns *Cythea* and *Dicksonia antarctica* (24%) in the upper part of the diagram suggests that the climate was at least as warm as the post-glacial climatic optimum, when the value for tree-ferns in the King Valley was considerably lower (Fig 4.1).

The Regency Interglacial can be compared to zones 5 to 3 of the Langdon Interglacial (Colhoun *et al.* in press) which can be used as a model of the structural characteristics of interglacial climatic change and the successional establishment of temperate rainforest in western Tasmania. These zones are dominated by *Casuarina-Eucalyptus* (zone 5), *Phyllocladus-Casuarina* (zone 4), and *Phyllocladus-Casuarina-Bauera-Nothofagus* (zone 3). No such succession occurs in the Regency Interglacial. Prominent in the early stages of development of rainforest at Langdon river is the increasing abundance of *Phyllocladus* and near absence of *Nothofagus* as *Eucalyptus* and herbs decline. *Phyllocladus* continues to dominate as the amount of *Nothofagus* increases and *Casuarina* and *Eucalyptus* maintains stable values. *Phyllocladus* pollen in the Regency Interglacial maintains consistent values throughout the profile and never achieves the dominance that it does at Langdon River.

The greater abundance of *Cythea* in the Regency profile appears to have significance because in Holocene pollen diagrams (Macphail 1979) *Dicksonia* is the dominant tree-fern (E. A. Colhoun pers. comm. 1987). Towards the top of the Regency profile the appearance of the tree ferns and abundance of *Lagarostrobos franklinii* suggest the rainforest canopy was more open though the reason for it becoming open cannot be determined from the pollen analysis.

The Regency Interglacial, the oldest known interglacial flora from western Tasmania can be contrasted with what are regarded to be Oligocene, Pliocene, and early Pleistocene palynofloras. An interesting comparison is that the amount of extinct Nothofagus pollen found at several sites appears to decrease in the younger deposits (Table 8.1).

Table 8.1 Extinct *Nothofagus* species in Tasmanian rainforest palynofloras.

Site	% *brassi* type pollen
Pioneer[1]	47
Lemonthyme[2]	32
Gormanston[3]	13
Regatta Point[4]	trace
Regency	0

1. Hill and Macphail (1983)
2. Paterson et al. (1967)
3. Kiernan (1980)
4. Hill and Macphail (1985)

The abundance of extinct species in organic silts overlying the Lemonthyme tillite and the lack of extinct species in the Regency Interglacial except for traces of two extinct rainforest taxa *Gothanipollis perplexus* and *Quintinia psilatispora*, suggests that the Regency Interglacial and the interglacial at Lemonthyme are widely separated in time (Table 8.1) If the decreasing amount of extinct taxa recorded in rainforest environments in Tasmania is a pattern that is due to age, then the age of the pollen sequences are as they are ranked in Table 8.1. However as already noted both the causes and the time of extinction of the numerous rainforest taxa have yet to be established. What is apparent is that the change is almost certainly one that cannot be related to the Tertiary-Pleistocene boundary in the simple terms it has been in the past, nor can glaciation be invoked to explain the extinctions as the abundance of extinct taxa after the Lemonthyme Glaciation shows. The extinction of these species is almost certainly diachronous.

8.3.5 The Governor Glaciation.

Evidence for the Governor Glaciation is limited to two sites which show major weathering unconformities between the King and Governor Formations. Because sediments of the Governor Glaciation that were transported by the King Glacier have no depositional forms and crop out only beneath younger deposits, it is not possible to accurately reconstruct the extent of ice during the Governor advance. However, the large particle sizes of the sediments suggests the ice limit was close to Baxter Rivulet and that it would have been similar to that of the King advance (Fig. 2.6).

The Governor Glaciation consists of two advances. The outwash gravels of each advance are separated by organic-rich fluviatile sediments of the Baxter Interstadial. Pollen analysis of these sediments suggest there was a climatic amelioration between the advances. It is this inferred warming that is the basis for defining the Baxter Interstadial. The Baxter Interstadial records a wet heath (pollen zone BR 2) and a heath-herbland mosaic (BR 2, Fig. 4.3). The pollen assemblage is quite different from the palynofloras of other organic rich deposits in its very high values of *Casuarina* and very low values of *Eucalyptus*. The same contrast has been noted between an interglacial pollen/vegetation succession at Langdon River which has a sequence of *Casuarina monilifera, Phyllocladus aspleniifolius* and *N. cunninghamii* (Colhoun *et al.* 1988 in press) and the Holocene succession of *Eucalyptus, Phyllocladus, Nothofagus and Eucryphia* (Macphail 1979; Colhoun and van de Geer 1986).This pollen analysis indicates a cool non-forested interstadial type climate, an interpretation that the overlying glacigenic deposits support. However, it is possible, though not determinable from the section that the two pollen zones BR2 and BR1 may represent the final part of an interglacial sequence as climate was deteriorating.

Assuming that the interstadial interpretation is correct then it is likely that mean annual temperature at the time of formation of the Baxter Interstadial sediments would have been considerably lower than present. Since the composition of the palynoflora suggests a treeline close to the altitude of Baxter Rivulet (c180 m), a temperature depression of 4-5°C is implied because present treeline occurs at about 1000 m.

8.3.6 The Henty Glaciation.

Deposits of the Henty Glaciation are the most common glacial deposits in the King Valley. The glaciation is represented by three advances the latest two of which (the Bull Rivulet and Blackwood advances) may be separated by a very short period of time. Separation of the King and Blackwood advances is based on the presence of over 40 m of non-glacial lake sediments between them.

At the maximum extent of the Henty Glaciation the King Glacier was over 300 m thick and almost as extensive as during the Thureau advance. Geological processes at this time, though dominated by glacial action also included widespread periglacial and paraglacial action which led to the deposition of thick angular screes that are interbedded with and overlie Henty Glaciation sediments. On the eastern slopes of Mt. Owen, parts of these screes became solifluction lobes which extend down onto the floor of the valley and across the Blackwood Formation end moraine. The widespread nature of periglacial deposits suggest that during and after ice retreat from the extensive Blackwood and King advances, the climate remained cold for quite some time before the slopes became vegetated.

Several NE-SW trending lineations on the Blackwood Formation aggradation surface indicate that basement faulting has occurred after deposition. There is little evidence of Quaternary faulting elsewhere in Tasmania.

The Nelson River Interstadial is defined by the absence of diamictons in the 40+m of lake sediments rather than the presence of an interstadial flora. The lack of dropstones and glacigenic sediments in these sediments suggests the sediments accumulated at the bottom of a lake that was beyond glacial influences and that at least the middle part of the King Valley was free of ice at this time. On this basis the sediments must be regarded as of interstadial character because they represent a comparatively long lived non glacial period in a glacial event, even though the climate may have been cold.

8.3.7 The Langdon Interglacial.

The Langdon Interglacial is known from pollen and plant macrofossils that occur outside the limits of the Last Glaciation and overlie Henty Glaciation deposits at Langdon River, about 12 km west of the King Valley. The Langdon Interglacial records the successional development of temperate rainforest (Colhoun *et al.* 1988 in press).

No deposits of the Langdon Interglacial are known in the King Valley. However, organic sediments at Smelter Creek (G.R. 865289) record a temperate rainforest flora which may have been deposited during the last interglacial (G. van de Geer and E. A. Colhoun pers. comm. 1987). However these sediments are not dated and they may be older than the Langdon Interglacial.

8.3.8 The Margaret Glaciation.

In the King Valley, deposits of the Margaret Glaciation are limited to the upper part of the valley near Dante Rivulet. Prior to this thesis the Last Glaciation was thought to be relatively simple with one advance that occurred about 18,000 yrs. ago (Kiernan 1983b; Colhoun 1985b). However, recognition of the Chamouni Formation as being of Margaret Glaciation age suggests the stratigraphy of the Margaret Glaciation is more complex.

The main issue is that till of the Chamouni Formation of Margaret Glaciation age lie up to 4 km outside the limits of the Dante advance. This implies that the climate was cool enough for glaciers to form during the Margaret Glaciation prior to 21 ka. B.P. Although the evidence for an earlier ice advance during the Margaret Glaciation is compelling for the King Valley, no similar advance has yet been recorded elsewhere in The West Coast Range.

Pollen from the silts that conformably rest beneath the Chamouni Formation record an alpine-subalpine assemblage that is consistent with the onset of the Chamouni advance (Table 4.3).

A palynoflora from Newall Creek records an early Last Glaciation age flora. This deposit records cool, open Eucalypt forest that existed prior to what appears to be the Dante advance (van de Geer *et al.* 1988 in press).

Pollen analysis of the alpine humus palaeosol that underlies the Dante Formation outwash gravel (Fig 4.73) suggests that a cold climate herbfield-bog mosaic existed prior to the Dante advance (Gibson *et al.* 1987). A similar palynoflora is recorded in the lower King Valley which was not glaciated during this time (Fig. 8.2). The late glacial part of this flora, preserved in a filled sinkhole, is dominated by subalpine and alpine trees and shrubs (Fig 8.2).

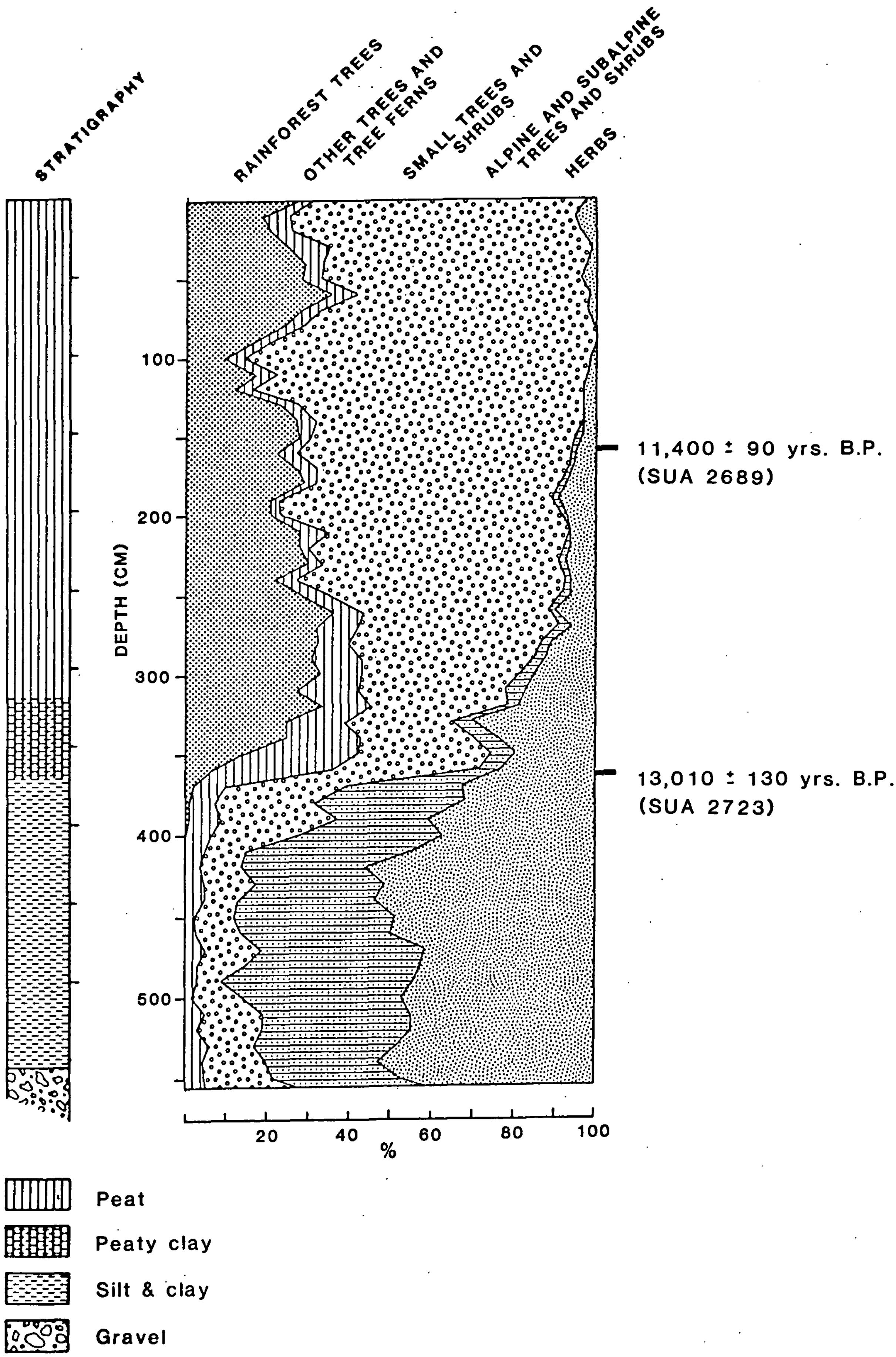


Fig. 8.2. Summary diagram from a late Pleistocene-Holocene silt-peat sequence in the lower King Valley (Pollen analysis by G. van de Geer).

Ice was absent from the floor of the valley during most of the Margaret Glaciation. The major geomorphological changes were caused by outwash streams and periglacial processes acting on deforested slopes. The degree to which periglacial processes modified the landscape at this time is difficult to assess because the deposits cannot be distinguished from older periglacial slope mantles. However, most of the periglacial deposits in the King Valley are believed to have formed as paraglacial (Church and Ryder 1972) slope deposits immediately after the Henty Glaciation, and during it in unglaciated areas. One of the major geomorphic effects of the Margaret Glaciation was the entrenchment of the King River into thick middle Pleistocene glacial deposits that cover much of the valley floor. From the middle part of the valley to the entrance of the gorge through the West Coast Range, the King Valley forms a relatively steep-sided gorge incised up to 40 m into aggradation surfaces of the King and Blackwood formation. The King River is now rock-floored for most of its length. The eastern bank consists of rock and the western of Quaternary deposits.

The transition from the Margaret Glaciation to the Holocene appears to have been abrupt (Fig. 8.2) and occurred around 13ka. BP. The date for the filled sinkhole (13,010 ± 130 yrs. B.P. SUA 2723) is confirmed by the presence of tree trunks in Holocene gravels in the upper King Valley dated at 12,250 ± 90 yrs. B.P. (SUA 2415), suggesting that large trees were in the King valley at this time and that ice was absent. These minimum dates for the deglaciation of the King valley are slightly older than most other sites for the West Coast range area eg Poets Hill (11,420 ± 700 yrs. BP Gak 1826, Davies 1974) and Tullabardine Dam (11,660 ± 150 yrs. BP SUA 1044, Colhoun and van de Geer 1986).

8.3.9 The Holocene.

There has been little geomorphic activity in the King Valley during the Holocene. In the upper valley, the King River has reworked older glacigenic sediments and formed low-lying deposits. Similar reworking occurs in isolated low lying areas in the lower King Valley and extensive areas in the Nelson River.

Excavation of a sediment filled depression in the lower King Valley shows there was inorganic sedimentation from 5.5 m to 3.7 m, with organic sedimentation occurring from 3.7 m to the surface (Fig. 8.2). Pollen analysis of these sediments suggests the post-glacial climatic optimum occurred from about 9-6 ka. BP. A probable change to cooler conditions after 6 ka is marked by an increase in the number of small trees and shrubs and a decline in the amount of pollen from rainforest trees. The inferred age of the post-glacial climatic optimum is similar to that noted by Macphail (1979) as occurring from 8-5 ka. B.P. on the Central Plateau, and the inferred cooling is similar to that suggested for Mt Field (Macphail 1979).

Formation of Holocene peat has occurred throughout western Tasmania and in the King Valley. It forms blankets from 0.4 to 1.4 m deep over most poorly drained surfaces with slopes up to 20°. Similar peat blankets have been described in Europe and their formation has been attributed to the transition from glacial climates to post-glacial optimal warmth at about 6000 yrs. B.P. (Haftsen and Solem 1976). The same climatic inferences about formation may not apply to Tasmania. However, the development of blanket peats in the King Valley probably results from the change in the relationships between precipitation and temperature that occurred during and after deglaciation.

Two sections in the study area at Newall Creek and in the middle King Valley near the Thureau Hills indicate that some slope instability has occurred during the Holocene. Although several others have inferred that slopes were unstable in the Holocene (Wasson 1977, Caine 1978) it is not possible to determine if this was due to changing climate or local geomorphological effects.

8.4 Correlation.

Correlation of sediments and rocks aims to demonstrate correspondence in character and/or stratigraphic position (Hedberg 1976). This discussion concerns temporal correlation.

In order to demonstrate temporal correspondence the age of the correlated sediment bodies must be known. When correlation precedes dating it is difficult to argue about contemporaneity let alone about age differences (Vita-Finzi 1973). Having recognised this, and considering the lack of securely dated sediments beyond the range of ^{14}C dating in Tasmania, any attempt at correlation of glacial events requires several assumptions. The most important of these is that the major glacial events were broadly synchronous. However, assumptions about the contemporaneity of glacial events between Tasmania and similar regions in New Zealand and southern South America achieve little when one of the purposes of international correlation is to determine if glacial events were contemporaneous.

8.4.1 Regional Correlation.

To attempt a correlation of glacial events within Tasmania requires an assessment and comparison of the chronologies of other studies. Such a comparison is beset by several problems that include a lack of relative dating data, absence of absolute dates for deposits older than the late part of the Last Glaciation, and lack of a consistent approach to field mapping and stratigraphic classification.

To produce a correlation from the data cited in other studies requires assessing how the interpretations were made and understanding the relevance of the events deduced to the chronology of the King Valley.

A suggested correlation of the major glacial sequences in Tasmania is summarised in Table 8.2. The correlations have been made on the basis of radiocarbon dating for the late Last Glaciation and analysis of weathering rind data from Jurassic dolerite clasts

Of the studies that are assessed here the writer is only familiar with the field areas of the recent studies of Bowden (1974), Sansom (1978), Kiernan (1980), and Augustinus (1982).

TABLE 8.2 Correlation of Quaternary glacial sequences in Tasmania.

Glaciations	Interglacials	King Valley (this thesis)	Central West Coast Range (Kiernan 1980)	Pieman Valley (Augistinus 1982)	Ben Lomond (Caine 1983)	Lake St Clair, central plateau (Kiernan 1985)	Mersey Valley (Hannan and Colhoun 1987)
Margaret		Dante	Dante	—	Cirque	Cynthia Bay	Rowallen
		Chamouni				—	—
	Langdon	—	—	—	—	—	—
Henty		Bull Rivulet Blackwood King	Comstock	? Boco 2 ? Boco 1	Plateau	Beehive Powers Creek Clarence	Arm
Governor		Governor/ Fish		? Boco 1	—	—	—
		Traveller			—	—	—
	Regency	Regency	—	—	—	—	—
Linda		Thureau	Linda	? Bulgobac ? Que	—	Stonehaven	Croesus
	Lemonthyme interglacial	—	—	—	—	—	—

Though there were several early studies that attempted to describe the glaciation of the West Coast Range area (see Banks *et al.* 1987), and multiple glaciation was suggested by Lewis (1945), the more recent studies of the area (Ahmad *et al.* 1959; Derbyshire 1963; Read 1963) suggested that most of the data previously observed could be explained as the product of a single glacial cycle. Since none of these studies were based on detailed stratigraphic analyses and did not have the availability of modern dating methods they do not contain data relevant to the correlation of glacial events in the King Valley.

Stratigraphic studies of the glacial sediments of the area began in the 1970's with Bowden's (1974) and Sansom's (1978) reconnaissance studies of glaciation in parts of western Tasmania. Although they found evidence for multiple glaciation, they do not have a consistent approach to stratigraphic classification or to field mapping.

A later (1980) thesis by Kiernan and two papers arising from it (Kiernan 1983a and 1983b) were early studies that addressed questions of the stratigraphy and antiquity of Tasmania's glacial sediments. The area he studied, the central part of the West Coast Range, extends into the King Valley.

Kiernans work established a chronology of three glacial stages, the Dante (equivalent to the Margaret), Comstock (equivalent to the Henty), and Linda Glaciations, which have remained essentially unchanged since they were initially defined. Differentiation of the stages was based on differences in weathering rind thickness on Jurassic dolerite clasts and radiocarbon dating of the late Last Glaciation. Large differences in weathering rind thicknesses were interpreted as representing interglacial warming.

The nomenclature of the glaciations has a complex history. The term "Margaret" for the Last Glaciation has been preserved from the early work of Lewis (1945). The term "Henty" comes from the recognition by Banks *et al.* (1977) that the till at Henty bridge predated the Last

Glaciation. The till compares with that described by Lewis at Yolande bridge where it cannot be dated. The terms "Dante" and "Comstock" were used by Kiernan (1980) for Last Glaciation and pre Last Glaciation drift on the eastern side of the West Coast Range. Both Dante and Comstock have not been used in this thesis because it appears that they are the equivalents of Margaret and Henty. The term Linda was adopted because the deposits in the Linda Valley were originally described by Lewis as being of the same age as those at Malanna which have since been shown to be non-glacial (Banks and Ahmad 1959). Since the Linda Valley was evidence for the oldest glaciation in Tasmania the name was selected to recognise its priority (E. A. Colhoun pers. comm. 1988).

The Dante "Glaciation" (Kiernan 1983a), is clearly the equivalent of the last or Margaret Glaciation (Table 8.2). The Dante Formation is now regarded as a late Last Glaciation advance of the King Glacier and has formation rather than glaciation status. The Chamouni Formation, a last glaciation advance of the King Glacier up to 4 km beyond the limits of the Dante sediments does not correlate with any sediments identified by Kiernan.

The Comstock Glaciation defined by Kiernan includes deposits defined by this study as Chamouni, Bull, Blackwood and King formations. Although Kiernan (1980) subdivided deposits mapped as Comstock glaciation into four parts (Comstock 2, Comstock 3, Linda 4, Sedgwick 1) the divisions appear to be based on altitude and geographic position rather than on stratigraphic position. The divisions cannot therefore be related to the glacial advances identified in this study. The name Comstock should be dropped in favour of Henty because all known evidence suggests that they are approximately the same age.

The Linda Glaciation was originally identified as a glacial stage that is separated from the Henty (Comstock) glaciation on the basis of extreme weathering of Jurassic dolerite clasts (Kiernan 1983b). Subsequently it has been shown that the Linda Glaciation lake sediments have a reversed detrital remanent magnetisation (Barbetti and Colhoun 1988 in press and M. Pollington pers. comm. 1986). Essentially the same criteria have been used for its identification in this study. In addition, the recognition of the organic sediments of the Regency Formation demonstrate that the Linda Glaciation was followed by interglacial warming and the development of temperate rainforest. The subdivisions that Kiernan (1980) made within the Linda Glaciation deposits (Linda 1 and Linda 2) cannot at present be demonstrated to belong to different glacial advances.

Augustinus (1982) and Augustinus and Colhoun (1987) analysed the glacial stratigraphy of the Pieman Valley in the northern part of the West Coast Range about 35 km northwest of the King Valley (Fig. 1.2). Although their study initially appears of a similar nature ,though more comprehensive than Kiernan's (1980 and 1983), it contains several major flaws and inconsistencies. Glacial deposits were differentiated as "drift sheets" that "probably represented first order glacial events". Although the meaning of a first order glacial event is unclear, it is used in a manner that suggests that deposits were divided into glacial stages that were differentiated primarily by their weathering characteristics. By restricting mapping to that of glaciations, Augustinus appears to ignore the possibility that glaciations contain multiple glacial advance and treats all observed differences in weathering characteristics as inferring interglacial warming. This particular problem is marked by differences between the thesis, which takes a conservative approach and identifies two glacial stages, and the publication (Augustinus and Colhoun 1987) which identifies four glacial stages with the same evidence. The four glacial stages are Boco 2, Boco 1, Bulgobac and Que Glaciations.

Familiarity with the Pieman Valley suggest that there are several dubious field relationships and inconsistencies in this study. These occur in the collection and use of relative dating data, recognition of the Que Glaciation and separation of the Boco 1 and 2 glaciations.

The main relative dating methods used by Augustinus were water absorption and specific gravity of Cambrian volcanic clasts and weathering rind thickness on Jurassic dolerite clasts. While water absorption and specific gravity data for the Boco 1, Boco 2 and Bulgobac sediments are different there are inconsistencies in the collection and interpretation of the data. One of the problems with using data of this type is that its collection should be restricted to one lithology to minimise effects other than time in the weathering process. Augustinus collected bulk samples of Cambrian volcanic rocks in an attempt to minimise the error from lithology. However, his own data on the lithology of the tills of the Pieman Valley demonstrates that there is a wide variety of Cambrian volcanic lithologies between sections within the same drift sheets (Tables 3.1, 4.1 and 5.1 in Augustinus 1982). Observations on the weathering of Cambrian volcanic clasts in tills in the King Valley suggest that different lithologies have very different weathering characteristics. Specifically, poorly indurated tuffs weather considerably faster than other more siliceous and more indurated lithologies. Because of the wide variety of lithologies in the Mt. Read Volcanics the results reported by Augustinus may mean little in terms of age and may be related to the source rock lithology of the glacial sediments.

Evidence that the Boco 1 sediments were overridden by the Boco 2 "glaciation" is not consistent with the description of the glacial landforms of the Boco 1 deposits. They are described as subdued when they should be eroded or buried if indeed they were overridden. Together the evidence suggests that the separation of the Boco 1 and Boco 2 sediments and their classification as different glaciations is not supported by the data. Augustinus (1982) also claims that there is an order of magnitude difference between the weathering data for the Boco 1 and Bulgobac sediments. However, the data quoted to support this is based on a comparison between mean weathering rind thickness of the Boco 1 sediments and the maximum thickness of weathering rinds in the Bulgobac sediments.

Recognition of the Que Glaciation is based on the occurrence of erratics beyond the limits of continuous Bulgobac drift, the altitude of deposits and by differences in clay mineralogy. None of the evidence suggests that the Que deposits are any different from those of the Bulgobac drift.

Because of these difficulties in Augustinus' data, correlation of some of it with other glacial sequences in Tasmania must remain dubious. For this reason the suggested correlations with the King Valley sequence are suffixed with question marks (Table 8.2).

Caines' (1983) study of the glaciation of the Ben Lomond massif suggested that there were deposits that related to two separate glaciations which he named the Cirque and Plateau. These are correlated with other glacial sequences in Tasmania on the basis of a count back and weathering characteristics of Jurassic dolerite (Table 8.2).

Kiernan's (1985) study of the glacial deposits of the Central Plateau is the most extensive study of glaciation in Tasmania. The size and lack of exposure in the study area dictated an approach that synthesised the mainly morphological evidence from several catchments. Field mapping of Kiernan's study was basically morphostratigraphic, that is it used moraines and outwash surfaces as primary mapping units in a similar manner as was used in this study. Relative dating of the deposits was primarily based on weathering rinds on Jurassic dolerite, but also included a wide range of criteria related to post-depositional modification of landforms and sediments.

Kiernan has recognised three weathering zones that are surficial rock units delimited from others by their weathering characteristics (Boyer and Pheasant 1974). The purpose of defining weathering zones is unclear, but appears to be to avoid the misuse of stratigraphic conventions in which evidence of interglacials is required before sediments can be classified as belonging to different glaciations. Weathering zone 1 includes deposits of the Stonehaven drift which Kiernan suggests is a correlate of the Linda Glaciation. Weathering zone 2 includes the Beehive, Powers Creek and Clarence sediments which are collectively known as the Butlers Gorge Complex and have not yet been correlated with the deposits of the West Coast region. Weathering zone 3 consists of the little weathered Cynthia Bay sediments.

Although Kiernan applies three weathering rate curves to estimate the ages of these deposits, the estimates differ widely. They suggest the the comparison of weathering characteristics of widely dispersed sites is not sufficiently accurate to allow a true appreciation of the relative age of all deposits.

The weathering of Cynthia Bay sediments suggests they are a correlate of the Margaret Glaciation. On the basis of counting backwards and weathering rinds, the Butlers Gorge Complex is a correlate of the Henty and/or Governor glaciations. Any more detailed correlation of individual advances is not possible without being able to obtain absolute ages of the sediments.

The advanced state of weathering of the Stonehaven sediments with respect to that of the Butlers Gorge Complex suggest that it is about the same age as Thureau Formation sediments in the King Valley (Table 8.2).

Correlation of glacial sediments within Tasmania is based on comparisons of the degree of weathering of Jurassic dolerite and matching ages by counting backwards from the late Margaret Glaciation advance, which is the only dated glacial advance in Tasmania.

Another difficulty in correlation that is based on weathering indices occurs when chronologies become more refined is that the differences in weathering. Small numerical differences in weathering cannot be accepted as a means of long distance correlation or for differentiating glacial advances. It is for this reason that the glacial stratigraphy of Tasmania has not become significantly more detailed or refined since the initial application of weathering as a relative dating method by Kiernan (1980, 1983b). This has occurred despite the subsequent intensive effort put into relative dating of deposits by weathering indices (Augustinus 1982; Kiernan 1985; Colhoun 1985b; Augustinus and Colhoun 1987).

Regional correlation based on weathering is bound to lead to errors and inconsistencies yet it is the only method available at present. Until the sediments are dated, it will not be possible to prove temporal correspondence of glacial events and suggestions about correlations will remain speculative.

On the basis of the attempts at correlation made here and consideration of other stratigraphic studies of glacial deposits in Tasmania several conclusions can be drawn.

1. Four separate glacial stages the Margaret, Henty, Governor and Linda glaciations affected the West Coast Range of Tasmania. Evidence for an older glaciation (the Lemonthyme Glaciation) is limited to an area in the Forth Valley (Paterson *et al.* 1967).

2. Two interglacial stages, the Langdon and Regency interglacials are known from the same area. In the King Valley only deposits of the Regency Interglacial are known, the existence of two other interglacials is inferred.

3. The Dante Formation represents a late Margaret Glaciation ice advance but it is not the oldest or the most extensive ice advance during the Margaret Glacial Stage.

4. The Chamouni Formation represents an ice advance that pre-dates the Dante Formation but appears to be considerably younger than deposits of the Henty Glacial Stage. The balance of evidence suggests it represents an ice advance in the early part of the Last Glacial Stage.

5. The Bull, Blackwood and King formations all represent ice advances that are $>^{14}C$ dating and are younger than the interstadial sediments at Baxter Rivulet which have a normal detrital remanent magnetisation.

6. The Governor, Fish and Traveller formations represent ice advances of the King and Jukes glaciers, they are $>^{14}C$ dating and are within the Brunhes magnetochron. The overlying King Formation is separated by a weathering unconformity that clearly suggests they are deposits of a separate glaciation named the Governor Glacial Stage.

7. Deposits of the Thureau Formation overlie Tertiary deposits and have a reversed detrital remanent magnetisation. They are overlain by organic deposits of the Regency Interglacial which records the successional development of temperate rainforest.

8.4.2 Correlation with glacial events in the Southern Hemisphere middle latitudes.

Correlation of Tasmania's glacial sequence with chronologies from South America and New Zealand is hampered by three considerations:

1. lack of absolute dates on glacial sediments older than Last Glaciation age;

2. correlation of glacial sediments in Tasmania is primarily based on weathering rind thickness of Jurassic dolerite and matching events on the basis of a simple count back assuming synchroneity of events. The weathering results are not comparable to weathering of glacial deposits in New Zealand or South America because glacial deposits in these areas do not contain Jurassic dolerite and have different climates that control weathering rates;

3. synchroneity of glacial events and land ice development in the Southern Hemisphere cannot be assumed, it must be demonstrated through the dating of deposits in each area.

Several correlations of Tasmanian glacial sequences with those of South America have been suggested. Three of these are summarised by Table 8.3.

Kiernan (1980, 1983b) suggested that the Dante Glaciation as it was then called, was a correlate of the Kumara 2^2 advance of the Otira glaciation defined by Suggate (1965a), (Table 8.3) and the Late Llanquihue Glaciation in South America (Mercer 1976). In the later paper Kiernan also suggested the Comstock glaciation which he was now calling the Henty Glaciation, may be a correlate of the Kumara 2^1 advance of the Otira Glaciation and/or the Waimea or Waimaunga glaciations (Table 8.3). Although in the 1980 study Kiernan did not attempt to correlate the Linda Glaciation, in 1983 he inferred a correlation between it and the Porika Glaciation (Table 8.3), though he also suggested the Linda Glaciation may be a great deal older.

Table 8.3 Suggested correlations of Tasmanian glaciations with the glacial sequence of New Zealand.

<table>
<tr><th colspan="2">Suggate (1965a)*</th><th>Kiernan (1980)</th><th>Kiernan (1983b)</th><th>Augustinus (1982)</th><th>Colhoun (1985b)</th><th>Kiernan (1985)</th></tr>
<tr><td rowspan="3">Otira Glaciation</td><td>Kumara 3 (Moana)</td><td>—</td><td></td><td>—</td><td>—</td><td></td></tr>
<tr><td>Kumara 2[2] (Larrikins)</td><td>Dante Glaciation</td><td>Dante Glaciation</td><td>—</td><td>Margaret Glaciation maximum</td><td>Cynthia Bay</td></tr>
<tr><td>Kumara 2[1] (Loopline)</td><td>—</td><td>?</td><td>—</td><td>—</td><td rowspan="3">Bultlers Gorge Complex</td></tr>
<tr><td colspan="2">Waimea Glaciation</td><td>Comstock Glaciation</td><td rowspan="2">Henty Glaciation</td><td>Boco 2 Glaciation</td><td>—</td></tr>
<tr><td colspan="2">Waimaunga Glaciation</td><td>—</td><td>Boco 1 Glaciation</td><td>—</td></tr>
<tr><td colspan="2">Porika Glaciation</td><td>—</td><td>Linda</td><td>Bulgobac Glaciation</td><td>—</td><td>Stonehaven</td></tr>
<tr><td colspan="2">Ross Glaciation</td><td>—</td><td>—</td><td>Que Glaciation</td><td></td><td>—</td></tr>
</table>

* Revised nomenclature of Suggate (1985a) in parentheses.

Colhoun (1985a) suggested that the maximum of the Margaret Glaciation, as dated at Dante Rivulet correlates with the Kumara 2[2] advance in Westland, the late Llanquihue glaciation in South America, and the peak of stage 2 of the oxygen isotope record from Indian Ocean cores. He regarded this as evidence which demonstrates the synchroneity of the Last Glaciation Maximum in west coast climates of the Southern Hemisphere. Although Colhoun makes no further correlations with the New Zealand glacial sequence, he suggests correlations with the South American on the basis of similar weathering rind thicknesses. Specifically he suggests that Margaret Glaciation deposits, as well as correlating with the late Llanquihue deposits, may also correlate with Llanquihue 1 deposits described by Mercer (1976), and Henty Glaciation deposits may correlate with Santa Maria, Rio Llico and Caracol drifts on the basis of similarity in the thickness of weathering rinds on volcanic clasts in South America and Jurassic dolerite in Tasmania.

Kiernan (1985) suggested correlations between Cynthia Bay sediments and the Kumara 2[2] advance in New Zealand and the maximum of the Llanquihue in South America. Assuming broadly similar hemispheric responses to climatic change, he correlated the Butlers Gorge Complex with the Waimea and or Waimaunga glaciations in Westland (Table 8.3) and pre Llanquihue glacial deposits in South America, and Stonehaven sediments with the Porika glaciation in New Zealand which was believed at that time to be about 750,000 yrs. old (Mildenhall and Suggate (1981).

Since these correlations were suggested the status of the New Zealand glacial chronology has changed significantly and now both the Porika and Ross glaciations are thought to be of Pliocene age (Suggate (1985b). The change in the N.Z. chronology disrupts many of the correlations that have been suggested (Table 8.3). It also raises the question of whether correlations of the glacial stratigraphy of Tasmania should be adjusted to accommodate changes in its supposed international correlates. To suggest that it should is ridiculous. This question highlights the dangers of correlating without dating, and emphasises that the fundamental meaning of correlation is to demonstrate temporal correspondence. Correlations suggested in Table 8.3 other than that of the dated late Margaret Glaciation maximum assume synchroneity in the responses to climatic change and development of land ice in the Southern Hemisphere.

Having criticized the previous attempts to correlate glacial events in Tasmania with those in South America and New Zealand, it is now possible to comment on what correlations are possible. They are few because of the serious limitations of dating methods as discussed previously. On the basis of ^{14}C dates the Dante Formation <(18,800 ± 550 yrs. BP ANU 2533) appears to be a correlate of the the Kumara 2^2 advance in New Zealand (23,300-18,000 yrs. BP, Suggate and Moar 1970) (now the Larrikins advance, Suggate 1985a) and the late Llanquihue glaciation in South America (19,400-17,300 yrs. BP, Mercer 1976). There is no basis for correlating the Henty, Governor and Linda glaciations with any glaciation in New Zealand or South America. However, if the interpretation of the palaeomagnetic boundary crossed by the Thureau Formation proves to be correct, then the Linda Glaciation has no known correlate with the glacial chronology of New Zealand as summarised by Suggate (1985b). Until the interpretation is proven this suggestion is mere speculation.

SUMMARY

- The King Valley gives clear evidence of multiple glaciation in Tasmania. It provides the clearest stratigraphic evidence yet obtained from the West Coast region which is the type area of Quaternary glaciation in Tasmania and Australia. The more complex stratigraphy developed from the King Valley can be used for comparison with other areas in Tasmania though direct correlation is still rather tenuous.

- Lack of suitable methods makes it impossible to date pre late Last Glaciation deposits adequately to provide reliable correlations with New Zealand and South America. The correlations that have been proposed are still fairly speculative.

REFERENCES

Ahmad, N., Bartlett, H. A. and Green, D. H. (1959) The glaciation of the King Valley, western Tasmania. *Pap. Proc. R. Soc. Tas.* 93, 11-17.

American Committee on Stratigraphic Nomenclature (1959) Report 6. Application of stratigraphic classification and nomenclature to the Quaternary. *Am. Assoc. Pet. Geol. Bull.* 43, 663-673.

Anderton, R. (1985) Clastic facies models and facies analysis. In *Sedimentology - recent developments and applied aspects.* (Ed. by P. J. Brenchley and B. J. P. Williams) p. 31-47. Blackwell, Oxford.

Andrews, J. T. (1971) *Techniques of till fabric analysis.* B.G.R.G. Technical bulletin No 6. 43pp.

Andrews, J. T. (1975) *Glacial systems.* Duxbury Press. Massachusetts 191pp.

Andrews, J. T. and Smith, D. I. (1970) Statistical anaylsis of till fabric : Methology, local and regional variability. *Quart. J. Geol. Soc. Lond.* 125, 503-542.

Ashley, G. M. (1975) Rhythmic sedimentation in glacial Lake Hitchcock,Massachusetts, Connecticut. In: *Glaciofluvial and Glaciolacustrine Sedimentation.* (Ed. by A. V. Jopling and B. C. McDonald.) p. 304-314. Soc. Econ.Palentologists and Mineralolgists Special Pub. No. 23.

Augustinus, P. (1982) *The glacial geomorphology of the middle Pieman - Bulgobac area.* BSc. (Hons) thesis, Dept. of Geography. University of Tasmania.

Augustinus, P. and Colhoun, E. A. (1987) Glacial history of the upper Pieman and Boco valleys, western Tasmania. *Aust. J. Earth. SciI.* 33, 181-191.

Banjeree, I. (1973) Part A: Sedimentology of Pleistocene glacial varves in Ontario, Canada. Part B: Nature of grainsize distribution of some Pleistocene glacial varves of Ontario, Canada. *Can. Geol Surv. Bull.* 226, 60p.

Banks, M. R. (1957) *The stratigraphy of Tasmanian limestones.* In: Limestones in Tasmania. (Ed. by T. D. Hughes.) Geol. Surv. Min. Res. No. 10 Department of Mines, Tasmania.

Banks, M. R. and Ahmad, N. (1959) Notes on the Cainozoic history of western Tasmania - the Malanna Glaciation. *Pap. Proc. R. Soc. Tas.* 93, 117-127.

Banks, M. R. and Ahmad, N. (1962) The Permian System in western Tasmania. *Pap. Proc. R. Soc. Tas.* 96, 1-18.

Banks, M. R.,Colhoun, E. A. and Chick, N. K. (1977) A reconnaissance of the geomorphology of central western Tasmania. In *Landscape and Man.* (Ed. by M. R. Banks and J. B. Kirkpatrick) p. 29-54. Royal Society of Tasmania. Hobart

Banks, M. R., Colhoun, E. A. and Hannan, D. (1987) Early discoveries of the effects of ice action in Australia. *J. Glaciol.* 33, 231-235.

Barbetti, M. and Colhoun, E. A. (1988, in press) Reversed magnetisation of glaciolacustrine sediments from western Tasmania. *Search*

Benett, K. D. (1986) Coherent slumping of early postglacial lake sediments at Lake Hall, Ontario, Canada. *Boreas* 15, 209-216.

Beu, A. G. and Edwards, A. R. (194) New Zealand Pleistocene and Late Pliocene glacio-eustatic cycles. *Palaeogeography Palaeoclimatology Palaeoecology* 46, 119-142.

Bloom, A. L. (1978) *Geomorphology. A systematic analysis of late Cenozoic landforms.* Prentice Hall, New Jersey. 520pp.

Boulton, G. S. (1967) The development of a complex supraglacial moraine at the margin of Sorbreen, NY Friesland, Vestspitsbergen*J. Glaciology* 6, 717-135.

Boulton, G. S. (1968) Flow tills and related deposits on some Vestspitsbergen glaciers. *J. Glaciology* 7, 391-412.

Boulton, G. S. (1970a) On the origins and transport of englacial debris in Savabard glaciers. *J. Glaciology* 9, 213-229.

Boulton, G. S. (1970b) On the deposition of subglacial and melt-out tills at the margins of certian Savabard glaciers. *J. GlaciologY* 9, 231-245.

Boulton, G. S. (1971) Till genesis and fabric in Savabard glaciers. In *Till : A Symposium.* (Ed. by R. P. Goldthwait) p. 41-72. Ohio State University Press Ohio.

Boulton, G. S. (1972a) Modern Arctic glaciers as depositional models for former ice sheets. *Quart. J. Geol. Soc. Lond.* 128, 361-393

Boulton, G. S. (1972b) The role of thermal regime in glacial sedimentation. In *Polar Geomorphology*. (Ed. by R. J. Price and D. E. Sugden.) p. 1-19. Institute of British Geographers Special Publication No 6.

Boulton, G. S. (1974) Process and patterns of glacial erosion. In *Glacial Morphology*. p 41-88. (Ed. by D. R. Coates) State University of New York. New York

Boulton, G. S. (1975) Processes and patterns of subglacial sedimentation. In *Ice Ages : Ancient and modern*. (Ed. by A. E.Wright and F. Moseley.) p. 7-42.Geological Journal special issue no 6.

Boulton, G. S. (1976) Till-its genesis and diagenesis. U.A.M. *Seria Geografia* 12, 65-80.

Boulton, G. S. (1978) Boulder shapes and grain size distributions as indicators of transport paths through a glacier and till genesis.*Sedimentology* 25, 773-799.

Boulton, G. S. and Eyles, N (1979) Sedimentation by valley Glaciers; a model and genetic classification. In *Moraines and Varves*. (Ed. by Ch. Schuchter) p. 11-23. A. A. Balkema. Rotterdam.

Boulton, G. S. and Jones, A. S. (1979) Stability of temperate ice capsand ice sheets resting on beds of deformable sediment. *J. Glaciology* 24, 29-43.

Boulton, G. S., and Paul, M. A. (1976) The influence of genetic processes on some geotechnical properties of glacial tills. *Q. J. Eng. Geol.* 9, 159-194.

Bowden, A. R. (1974) *The glacial geomorphology of the Tyndall Mountains*. BSc. (Hons.) thesis. Dept. of Geography. University of Tasmania.

Bowen, D. Q. (1978) *Quaternary Geology*. Pergamon Press, Oxford, 221pp.

Boyer, S. J. and Pheasant, D. R. (1974) Delimitation of weathering zones in the fiord area of eastern Baffin Island. *Geol. Soc. Am. Bull.* 85, 805-810.

Bradley, J. (1954) The geology of the West Coast Range of Tasmania, Part 1. *Pap. Proc. Roy. Soc. Tas.* 88, 193-243.

Broster, B. E. and Dreimanis, A. (1981) Deposition of multiple lodgement tills by competing glacial flows in a common ice sheet: Cranbrook, British Columbia. *Arctic and Alpine Res.* 13, 197-204.

Broster, B. E. and Hicock, S. R. (1985) Multiple flow support mechanisms and the development of inverse grading in subaquatic glacigenic debris flows. *Sedimentology* 32, 645-658.

Burke, R. M. and Birkland, P. W. (1979) Reevaluation of multi parameter relative dating techniques and their application to the glacial sequence along the eastern escaroment of the Sierra Nevada, California *Quaternary Research* 11, 21-51.

Caine, T. N. (1978) Late Holocene instability on hillslopes in northeastern Tasmania. *Search* 9, 298.

Caine, T. N. (1983) *The mountains of northeastern Tasmania: A study of alpine geomorphology.* A. A. Balkema, Rotterdam. 200pp.

Calver, C. R, Ballie, P. W., Everard, J. L., Williams, P. R., Forsyth, S. M., Turner, N. J. and Williams, E. (1987) Geological Atlas 1:50,000 series. Sheet 8013N Lyell.

Cheel, R. J. and Rust, B. R. (1986) A sequence of soft-sediment deformation (dewatering) structures in Late Quaternary subaqueous outwash near Ottawa, Canada. *Sedim. Geol.* 47, 77-93

Chinn, T. J. H. (1981) Use of rock weathering-rind thickness for Holocene absolute-age dating in New Zealand. *Arctic and Alpine Res.* 13, 33-45.

Church, M. and Gilbert, R. (1975) Proglacial fluvial and lacustrine environments. In *Glaciofluvial and Glaciolacustrine Sedimentation.* (Ed. by A. V. Jopling and B. C. McDonald.) p. 22-100. Soc. Econ. Palentologists and Mineralolgists Special Pub. No. 23.

Church, M. and Ryder J. M. (1972) Paraglacial sedimentation: A consideration of fluvial processes conditioned by glaciation. *Geol. Soc. Am. Bull.* 83, 3059-3072.

Clapperton, C. M. (1975) The debris content of surging glaciers in Svalbard and Iceland. *J. Glaciology* 24, 29-44.

Clague, J. J. (1975) Glacier flow patterns and the origin of late Wisconsin till in the southern Rocky Mountains, British Columbia. *Geol. Soc. Am. Bull.* 86, 721-731.

Clayton, L. and Moran, S. R. (1974) A glacial process-form model. In *Glacial Geomorphology.* (Ed. by D. R. Coates.) p. 89-120. State University, Binghampton, New York.

Cohen, J. M. (1979) Deltaic sedimentation in glacial Lake Blessington, County Wicklow, Ireland. In *Moraines and Varves. Origin/Genesis/Classification.* (Ed. by Ch. Schluchter.) p. 357-367. A. A. Balkema. Rotterdam.

Colhoun, E. A. (1975) A Quaternary climatic curve for Tasmania. Unpublished manuscript fron the Australian conference on climate and climatic change

Colhoun, E. A. (1974) A reconnaissance survey of the glaciation of Macquarie Island. *Pap. Proc. R. Soc. Tas.* 108, 1-19.

Colhoun, E. A. (1976) The glaciation of the lower Forth Valley, northwestern Tasmania. *Aust. Geog. Studies* 14, 83-102.

Colhoun, E. A. (1979) The Quaternary deposits of Blakes Opening and the middle Huon Valley, Tasmania. *Phil. Trans. R. Soc. Lond.* 1014, 371-395.

Colhoun, E. A. (1985a) Glaciations of the West Coast Range, Tasmania. *Quaternary Research* 24, 39-59.

Colhoun, E. A. (1985b) Pre-last glaciation maximum vegetation history at Henty Bridge, western Tasmania. *New Phytologist* 100, 681-690.

Colhoun, E. A. (1986) Radiocarbon dates for Tasmania 1956-1984. *Pap.Proc. R. Soc. Tasmania* 119, 39-64.

Colhoun, E. A. and van de Geer, G. (1986) Holocene to Middle Last Glaciation vegetation history at Tullabardine Dam, western Tasmania. *Proc. R. Soc. Lond.* B 229, 177-207.

Colhoun, E. A. and van de Geer, G. (1987) Vegetation history and climate before the last glaciation at Crotty, western Tasmania. *Pap.Proc. R. Soc. Tasmania* 121, 69-74.

Colhoun, E. A., van de Geer, G., Hill, R. S. and Bird, T. (1988 in press) Late Pleistocene organic deposits at Langdon River, western Tasmania. *New Phytologist.*

Colman, S. M. (1981) Rock weathering rates as functions of time. *Quaternary Research* 15, 250-264.

Colman, S. M., and Pierce, K. L. (1981) Weathering rinds on andesitic and basaltic stones as a Quaternary age indicator. *U. S. Geol. Surv. Prof. Paper* 1210.

Corbett, K. D. (1979) *Stratigraphy, correlation and evolution of the Mount Read Volcanics in the Queenstown, Jukes - Darwin and Mount Sedgwick areas.* Tasmania Department of Mines. Geological Survey Bull. 58.

Corbett, K. D. and Brown, A. V. (1976) Geological atlas 1:25,000 series sheet Sk-55/5, Queenstown.

Corbett, K. D. , Green G.R. & Williams, P. R. (1977) The geology of central western Tasmania. In: *Landscape and Man.* (Ed. by M. R. Banks and J. B. Kirkpatrick) p. 7-28. Roy. Soc Tas.

Corbett, K. D., Reid, K. O., Corbett, E. B., Green, G. R., Wells, K. and Shepard, N. W. (1974) The Mount Read Volcanics and Cambrian -Ordovician relationships at Queenstown, Tasmania. *J. Geol. Soc. Aust.* 21, 173-186.

Curtis, W. M. (1963) *The student's flors of Tasmania,* part 2. Government Printer, Hobart, Tasmania.

Curtis, W. M. (1967) *The student's flors of Tasmania,* part 3. Government Printer, Hobart, Tasmania.

Curtis, W. M. and Morris, D. I. (1975) *The student's flors of Tasmania,* 2nd. edn. part 1. Government Printer, Hobart, Tasmania.

David, T. W. E. (1926a) Summary of report of glacial phenomena committee.*Report of the Australasian Association of Advancement of Science* (1924) 17, 63-66.

David, T. W. E. (1926b) Pleistocene glaciation near Strahan. *Report of the Australasian Association of Advancement of Science* (1924) 17, 91-103.

Davies, J. L. (1959) High level erosion surfaces and landscape development in Tasmania. *Aust. Geog.* 7, 193-204.

Davies, J. L. (1962) Geomorphology and glaciation. In *The Geology of Tasmania*. J. Geol. Soc. Aust. 9, 243-250.

Davies, J. L. (1974) Geomorphology and quaternary environments. In *Biogeography and Ecology of Tasmania*. (ED. by W. D. Williams.) p. 17-27. D. W. Junk, The Hague.

Derbyshire, E. (1963) Glaciation of the Lake St. Clair district, West-Central Tasmania. *Aust. Geog.* 9, 97-110.

Derbyshire, E. (1966) Discussion of the Glacial Map of Tasmania. *Aust.J. Sci.* 29, 102-103.

Derbyshire, E. (1968) *Glacial map of west-central Tasmania*. Geological Survey Records 6, 1-4. Dept. of Mines, Tasmania.

Derbyshire, E (1972) Pleistocene glaciation in Tasmania: Review and speculations. *Aust. Geog. Studies* 10, 79-94.

Derbyshire, E., Banks, M. R., Davies, J. L. and Jennings, J. N. (1965) *Glacial map of Tasmania*. Roy. Soc. Tas. Spec. Publ. No. 2, 11p.

Derbyshire, E. and Peterson, J. A. (1971) On the status and correlation of Pleistocene glacial episodes in southeastern Australia. *Search* 2, 285-288.

Domack, E. W. and Lawson, D. E. (1985) Pebble fabric in an ice rafted diamicton. *J. Geol.* 93, 577-591.

Dowdeswell, J. A. (1982) Relative datind of Late Quaternary deposits using cluster and discriminant analysis, Audubon Range, Colorado Front Range. *Boreas* 11, 151-161.

Dowdeswell, J. A. and Sharp, M. J. (1986) Characterization of pebble fabrics in modern terrestrial glacigenic sediments. *Sedimentology* 33, 799-710.

Doweswell, J. A., Hambrey, M. J. and Wu, R. T. (1985) A comparison of clast fabric and shape in Late Precambrian and modern glacial sediments. *J. Sedim. Pet.* 55, 691-704.

Dreimanis, A. (1979) The problem of waterlain tills. In *Moraines and varves*. (Ed. by Ch. Schluchter) p. 167-177. A. A. Balkema Rotterdam

Dreimanis, A. (1981) Till: terminology, definition of till and relatedterms, genetic classification. INQUA Commission on the genesis and lithology of Quaternary deposits. Circular 23

Dreimanis, A. (1983) Quaternary glacial deposits: Implications for the interpretation of Proterozic glacial deposits. *Geol. Soc. Am. Memoir* 161, 299-307.

Dreimanis, A. (1987) Implications of the polygenetic orgin of till on palaeoenvironmental and stratigraphic interpretations. *INQUA abstracts* 1987

Dreimanis, A. and Schluchter Ch. (1985) Field criteria for the recognition of till or tillite. *Palaeogeography, Palaeoclimatology, Palaeoecology* 51, 7-14.

Drozdowski, E. (1977) Ablation till and related indicatory forms at the margins of Vestspitsbergen glaciers. *Boreas* 6, 107-114.

Dunn, E. J. (1893) *Remarks on the glaciation of Tasmania.* Evening Standard, Melbourne.

Dunn, E. J. (1894) Glaciation of the western highlands of tasmania. *Proc. Roy. Soc. Victoria* 6, (NS, 133-138).

Elson, J. A. (1961) *The geology of tills.* Proc. 14th. Can. Soil Mechanics Conference. Nat. Res. Canada Comm. Soil and Snow Mechanics Tech. Mem. 69

Ehlers, J. (1981) Problems of Saalian stratigraphy in the Hamburg area. *Meded. Rijks Geol. Dienst.* 34, 26-29.

Evenson, E. B., Dreimanis, A. and Newsome J. W. (1977) Subaquatic flow tills: A new interpretation for the genesis of some laminated deposits. *Boreas* 6, 115-133.

Eyles, N. (1979) Facies of supraglacial sedimentation on Icelandic andalpine temperate glaciers. *Can. J. Earth Sci.* 16, 1341-1361.

Eyles, N. and Clark. B. M. (1986) Significance of hummocky and swaly cross-stratification in Late Pleistocene lacustrine sediments of the Ontario basin, Canada. *Geology* 14, 679-682.

Eyles, N., Clark, B. M. and Clague, J. J. (1987) Coarse-grained sediment gravity flow facies in a large supraglacial lake. *Sedimentology* 34, 193-216.

Eyles, N., Eyles, C. H. and Miall, A. D. (1983) Lithofacies types and vertical profile models; an alternative approach to the description an environmental interpretation of glacial damict anddamicite sequences. *Sedimentology* 39, 393-410.

Eyles, N. and Menzies J. The subglacial landsystem. In *Glacial geology. An introduction for engineers and earth scientists.* (Ed. by N. Eyles) p. 19-70. Pergamon Press.

Eyles, N., and Paul, M. A. (1983) Landforms and sediments resulting from former periglacial climates. In:*Glacial geology. An introduction for engineers and earth scientists.* (Ed. by N. Eyles) Pergamon Press Oxford.

Eyles, N. and Rogerson, R. J. (1977) Glacier movement, ice structures and medial moraine form at a glacier confluence, Brendon Glacier, British Columbia, Canada. *Can. J. Earth Sci* 14, 2807-2816.

Eyles, N., Sladen, J. A. and Gilroy S. (1982) A depositional model for stratigraphic complexes and facies superimposition in lodgment tills. *Boreas* 11, 317-333.

Eyles, N. and Slatt, R. M. (1977) Ice marginal sedimentary, glaciotectonic and morphologic features of Pleistocene drift: An example from Newfoundland. *Quaternary Research* 8, 267-281.

Eynon, G. and Walker R. G. (1974) Facies relationships in Pleistocene outwash gravel in southern Ontario: a model for bar formation and growth in braided rivers. *Sedimentology* 21, 43-70.

Faegri, K. and Iversen, J. (1975) Textbook of pollen analysis, 3rd. edn. Munksgaard, Copenhagen.

Fleisher, P. J. (1986) Dead ice sinks and moats: environments of stagnant ice deposition. *Geology* 14, 39-42.

Folk, R. L., Andrews, P. B. and Lewis, D. W. (1970) Detrital sedimentary rock classification and nomenclature for use in New Zealand. *N. Z. J. Geol. Geophys.* 13, 43-70.

Francis, E. A. (1975) Glacial sediments: A selective review. In *Ice ages,ancient and modern.* (Ed. by A. E. Wright and F. Moseley) Geological Journal special issue no. 6

Gage, M. (1958) Late Pleistocene glaciations of the Waimakriri Valley, Canterbury, New Zealand. *N. Z. Geol. Geophys.* 1, 123-155.

Gage, M. (1960) On the definition, date and character of the Ross Glaciation, early Pleistocene, New Zealand. *Trans. Roy. Soc. N. Z.* 88, 631-638.

Gage, M. ed. (1977) *Stratigraphic nomenclature in the New Zealand Quaternary.* National Commettee for Quaternary Research, Royal Society of New Zealand. 25 pp.

Gentilli, J. (1972) *Australian Climatic Patterns.* Griffin Press, Adelaide.

Gibbons, A. B., Megeath, J. D. and Pierce, K. (1984) Probability of moraine survival in a succession of glacial advances. *Geology* 12, 327-330.

Gibson, N., Kiernan, K. W. and Macphail, M. K. (1987) A fossil plant bolster from the King River, Tasmania. *Pap. Proc. Roy. Soc. Tas.* 121, 35-42.

Gill, E. D. (1956) Radiocarbon dating for glacial varves in Tasmania. *Aust. J. Sci.* 19, 80.

Gill, E. D. (1962) Cenozoic. In *Geology of Tasmania.* J. Geol. Soc. Aust. 9, 107-354.

Gill, E. D. and Banks, M. R. (1950) Silurian and Devonian stratigraphy of the Zeehan area, Tasmania. *Pap. Proc. Roy. Soc. Tas.* (1949) 259-271.

Glen, J. W., Donner, J. J. and West, R. G. (1957) On the mechanism by which stones become oriented. *Am. J. Sci.* 255, 194-205.

Goldthwait, R. P. (1951) Development of end moraines in East- Central Baffin Island. *J. Geol.* 59, 567-577.

Gould, C. (1860) A report of the exploration of the western country by Mr. Gould. *Journals of the Legislative Council*, V, Government Printer, Hobart.

Grant-Taylor, T. L. and Rafter, T. A. (1963) New Zealand natural radiocarbon measurements. *Radiocarbon* 5, 118-162.

Gregory, J. W. (1904) A contribution to the glacial geology of Tasmania. *Quart. J. Geol. Soc. Lond.* 60, 37-52.

Gunn, B. M. (1964) Flow rates and secondary structures of Fox and Franz Josef Glaciers, New Zealand. *J. Glaciology* 5, 173-190.

Gustavson, T. C., Ashley, G. M. and Boothroyd, J. C. (1975) Depositional sequences in glaciolacustrine deltas. *Glaciofluvial and Glaciolacustrine Sedimentation.* (Ed. by A. V. Jopling and B. C. McDonald.) p. 264-280. Soc. Econ. Palentologists and Mineralolgists Special Pub. No. 23.

Haftsen, U. and Solem, T. (1976) Age, origin and palaeo-ecological evidence of blanket bogs in Norda-Trondelag, Norway. *Boreas* 5, 119-141.

Haines-Young, R. H. and Petch, J. R. (1983) Multiple working hypotheses, equifinality and the study of landforms. *Trans. Inst.Brit. Geogr.* 8, 458-466.

Haldorsen, S. and Shaw, J. (1982) The problem of recognising melt-out till. *Boreas* 11, 261-277.

Hamond, A. (1985) *Soils and glacial geomorphology of the central West Coast Range, Tasmania.* BSc. (Hons) thesis. Dept. of Geography. University of Tasmania.

Hannan, D. G. and Colhoun, E. A. (1987) Glacial stratigraphy of the upper Mersey Valley, Tasmania. *Aust. Geogr. Studies* 25, 36-46.

Hedberg, H. D. ed. (1976) *International Stratigraphic Guide.* Wiley. New york 200 pp.

Heusser, C. J. (1976) Palynology and depositional environment of the Rio Ignao nonglacialdeposit, province of Valdivia, Chile. *Quaternary Research* 6, 273-280.

Heusser, C. J. (1983) Quaternary palynology of Chile. In: *Quaternary of South America and the Antarctic Peninsula.* (Ed. by J. Rabassa.) p. 5-22. A. A. Balkema Rotterdam

Hill, R. S. and Macphail, M. (1983) Reconstruction of the Oligocene vegetation at Pioneer, northeast Tasmania. *Alcheringa* 7, 282-299.

Hill, R. S. and Macphail, M. (1985) A fossil flora from rafted Plio-Pleistocene mudstones at Regatta Point, western Tasmania. *Aust. J. Bot.* 33, 497-517.

Holmes, C. D. (1941) Till fabric. *Geol. Soc. Am. Bull.* 52, 1299-1354.

Hooke, R. L. (1968) Comments on "The formation of shear moraines : an example from south Victoria Land, Antarcitica". *Journal of Glaciology* 7, 351-352.

I.G.U.S. International Subconnission on Stratigraphic Classification and I.G.U.S./I.A.G.A. Subcommission on a Magnetic Polarity Time Scale. (1979) Magnetostratigraphic polarity units - A supplementary chapter in the I.S.S.C. International StratigraphicGuide. *Geology* 7, 578-583.

Imbrie, J. (1985) A theoretical framework for the Pleistocene ice ages. *J. Geol. Soc. Lond.* 142, 417-432.

Jennings, J. N. (1958) The Pleistocene glacial history of Tasmania. *J.Glaciol.* 3, 298-303.

Jennings, J. N. (1985) *Karst geomorphology*. Blackwell. 293 pp.

Jennings, J. N. and Ahmad, N. (1957) The legacy of an ice cap. *Aust. Geogr.* 7, 62-75.

Jennings, J. N. and Banks, M. R. (1958) The Pleistocene glacial history of Tasmania. *J. Glaciology* 3, 298-303.

Jenny, H. (1941) *Factors of soil formation*. McGraw Hill, New York.

Johnaston, R. M. (1894) The glacial epoch of Australiasia. *Proc. Roy. Soc. Tas.* (1893), 73-184.

Kamb, W. B. (1959) Ice petrofabric observations from Blue Glacier, Washington, in relation to theory and experiment. *Geophys. Res.* 64, 1891-1909.

Kamb, W. B. and La Chapelle (1964) Direct observation of the mechanism of glacier sliding over bedrock. *J. Glaciology* 5, 159-172.

Karrow, P. F. (1984) Discussion. Lithofacies types and vertical profile models; an alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences. *Sedimentology*31, 883-898.

Kiernan, K. W. (1980) *Pleistocene glaciation of of the central West Coast Range, Tasmania*. BSc. Hons. thesis, University of Tasmania.

Kiernan, K. W, (1982) Glaciation and karst in Tasmania: Review and speculations. *Helictite* 20, 11-16.

Kiernan, K. W. (1983a) Weathering evidence for an additional glacial stage in Tasmania. *Aust. Geog. Studies* 21, 197-220.

Kiernan, K. W. (1983b) Relationship of cave fills to glaciation in the Nelson River Valley, central western Tasmania. *Aust. Geogr* 15, 367-375.

Kiernan, K. W. (1985) *Late Cainozoic glaciation and mountain geomorphology in the central highlands of Australia.* PhD thesis, Dept. of Geography, University of Tasmania.

Kirkpatrick, J. B. (1977a) Native vegetation of the West Coast region of Tasmania. In *Landscape and Man.* (Ed. by M. R. Banks and J. B. Kirkpatrick) p. 55-80. Royal Society of Tasmania. Hobart.

Kirkpatrick, J. B. (1977b) The impact of man on the vegetation of the west coast region. In *Landscape and Man.* (Ed. by M. R. Banks and J. B. Kirkpatrick) p. 155-156. Royal Society of Tasmania. Hobart.

Kirkpatrick, J. B. (1984) Altitudinal and successional variation in thevegetation of the West Coast Range, Tasmania. *Aust. J. Ecol.* 9, 81-91.

Kirkpatrick, J. B. and Dickson, K. J. M. (1984) *Map of vegetation of Tasmania.* Forestry Commission, Hobart.

Kruger, J. and Marcussen, I. B. (1976) Lodgement till and flow till: A discussion. *Boreas* 5, 61-64.

Lawson, D. E. (1979a) *Sedimentological analysis of the western terminus region of the Matanuska Glacier, Alaska.* C.R.R.E.L. Report 79-9. U. S. Army Corps of Engineers. Hanover, New Hampshire.

Lawson, D. E. (1979b) A comparison of the pebble orientations in ice and deposits of the Matanuska Glacier, Alaska. *J. Geol.* 87, 629-645.

Lawson, D. E. (1981a) *Sedimentological characteristics and classification of depositional processes and deposits in the glacial environment.* C.R.R.E.L. Report 81-27. U. S. Army Corps of Engineers. Hanover, New Hampshire.

Lawson, D. E. (1981b) Distinguishing characteristics of diamictons at the margin of the Matanuska Glacier, Alaska. *Annals of Glaciology* 2, 78-84.

Lawson, D. E. (1982) Mobilization, movement and deposition of active subaerial debris flows, Matanuska Glacier, Alaska. *J. Geol.* 90, 279-300.

Lawson, D. E. (1987) Facies model for sedimentation along terrestrial glacier margins. *INQUA abstracts* 1987

Lewis, A. N. (1922) A preliminary sketch of the glacial remains preserved in the National Park of Tasmania. *Pap. Proc. Roy. Soc. Tas.* (1922), 32-39.

Lewis, A. N. (1926) Report on observations of glaciation in Tasmania since 1920. *Report of the Australasian Association of Advancement of Science* (1924) 17, 85-90.

Lewis, A. N. (1934) A correlation of the Tasmanian Pleistocene glacial epochs and deposits. *Pap. Proc. Roy. Soc. Tas.* (1933), 75-86.

Lewis, A. N. (1936) A record of Pleistocene varved shales from Tasmania. *Aust. Geog.* 2, 30.

Lewis, A. N. (1939) Notes on the Pleistocene glaciation, Mt. Field to Strahan. *Pap. Proc. Roy. Soc. Tas.* (1938), 161-173.

Lewis, A. N. (1945) Pleistocene glaciation in Tasmania. *Pap. Proc. Roy. Soc. Tas.* (1944), 41-56.

Lewis, D. W. (1984) *Practical sedimentology.* Hutchinson Ross, Pennsylvania. 227p.

Lindsay, J. F. (1968) The development of clast fabric in mud flows. *J. Sedim. Pet.* 38, 1242-1253.

Lindsay, J. F. (1970) Clast fabric of till and its development. *J. Sedim. Pet.* 40, 629-641.

Mabin, M. (1984) Late Pleistocene glacial sequence in the Lake Heron basin, mid Canterbury. *N. Z. J. Geol. Geophys.* 27, 191-202.

Macphail, M. K. (1979) Vegetation and climates in southern Tasmania since the Last Glaciation. *Quaternary Research* 11, 306-341.

Macphail, M. K. and Peterson, J. A. (1975) New deglaciation dates from Tasmania. *Search* 6, 127-129.

Marcussen, I. B. (1975) Distinguishing between lodgement till and flow-till in Wechselian deposits. *Boreas* 4, 111-123.

Mark, D. M. (1971) Rotational vector procedure for the analysis of till fabrics. *Geol. Soc. Am. Bull.* 82, 2661-2666.

Mark, D. M. (1973) Analysis of axial orientation data including till fabrics. *Geol. Soc. Am. Bull.* 84, 1369-1374.

Mark, D. M. (1974) On the interpretation of till fabrics. *Geology* 2, 101-104

Mark, D. M. and Andrews, J. T. (1975) A re-examination of till fabrics and origins of some "cross - valley" moraines on Baffin Island. *Geol. Foren. Fohr.* 97, 321-325.

Marksgraf, V., Bradbury, J. P. and Busby, J. R. (1986) Palaeoclimates in southwestern Tasmania during the last 13,000 years. *Palaios* 1, 368-380.

McDonald, B. C. & Shilts, W. W. (1975) Interpretation of faults in glaciofluvial sediments. In: *Glaciofluvial and glaciolacustrine sedimentation.* (Ed. by A. V. Jopling and B. C. McDonald) p. 123-131. S.E.P.M. Special Publication No 23.

McGregor, G. R. (1981) Weathering characteristics of Late Pleistocene tills. *N. Z. J. Geol. Geophys.* 24, 107-113.

Mercer, J. H. (1976) Glacial history of southernmost South America. *Quaternary Research* 6, 125-166.

Miall, A. D. (1978) Lithofacies types and vertical profile models in braided rivers: a summary. In *Fluvial Sedimentology.* (Ed. by A. D. Miall.) p. 597-604. Can. Soc. Petr. Geol. Memoir, 5.

Miall, A. D. (1982) Glaciofluvial transport and deposition. *In Glacial geology. An introduction for Engineers and Earth Scientists.* (Ed.by N. Eyles) p.168-183.Pergamon Press.

Mickelson, D. M. (1973) Nature and rate of basal till deposition in a stagnating ice mass. *Arctic and Alpine Res.* 5, 17-27.

Mildenhall, D. C. and Suggate, R. P. (1981) Palynology of the Tadmor Group (Late Miocene-Pliocene) and the Porika Formation (early Pleistocene), South Island, New Zealand. *N. Z. J. Geol. Geophys.* 24, 515-528.

Mills, H. H. (1977) Basal till fabrics of modern alpine glaciers. *Geol. Soc. Am. Bull.* 88, 824-828.

Moore, T. B. (1894) Discovery of glaciation in the vicinity of Mt. Tyndall in Tasmania. *Pap. Proc. Roy. Soc. Tas.* (1893), 147-149.

Nelson, C. S., Hendy, C. H., Jarrett, G. R. and Cuthbertson, A. M. (1985) Near-synchroneity of New Zealand alpine glaciation and northern hemisphere continental glaciation during the past 750k yr. *Nature* 318, 361-366.

Nye, J. F. (1952) The mechanics of glacier flow. *J. Glaciology* 2, 82-93.

Oyama, M. and Takehara, H. (1967) *Revised standard soil colour chart.*

Paterson, W. S. B. (1981) *The Physics of Glaciers.* 2nd. edn. Pergamon Press, Oxford. 380pp.

Paterson, S. J. (1965) Pleistocene drift in the Mersey and Forth valleys - probability of two glacial stages. *Pap. Proc. Roy. Soc. Tas.* 99, 115-151.

Paterson, S. J. (1966) Pleistocene deposits at Parangana damsite in the Mersey Valley. *Pap. Proc. Roy. Soc. Tas.* 100, 147-124.

Paterson, S. J., Duigan, S. L. and Joplin, G. A. (1967) Notes on the Pleistocene deposits at Lemonthyme Creek in the Forth Valley. *Pap. Proc. R. Soc. Tas.* 101, 221-226.

Paterson, S. J., Baynes, F. J. and Underwood, R. (1984) *King River Power Development consolidated geological report.* Vol. 1 Geology. Report 64 - 34 -3 Hydro Electric Commission of Tasmania.

Paul, M. A. (1983) The supraglacial landsystem. In *Glacial Geology. An introduction For Engineers and Earth Scientists.* (Ed. by N. Eyles) p. 71-90. Pergamon Press.

Peltoniemi, H. (1985) Till lithology and glacial transport in Kuhmo, eastern Finland. *Boreas* 14, 67-74.

Peterson, J. A. (1966) Glaciation of the Frenchmans Cap National Park. *Pap. Proc. R. Soc. Tas.* 100, 117-129.

Peterson, J. A. (1968) Cirque morphology and pleistocene ice formation conditions in southeastern Australia. *Aust. Geog. Studies* 6, 67-83.

Peterson, J. A. (1969) *The cirques of southeastern Australia: Studies in morphology and distribution; history and significance.* Phd. thesis, Monash University.

Pickrill, R. A. and Irwin, J. (1983) Sedimentation in a deep glacier fed lake- Lake Tekapo, New Zealand. *Sedimentology* 30, 63-75.

Porter, S. C. (1975) Weathering rinds as relative age criterion: Application to subdivision of glacial deposits in the Cascade Range. *Geology* 3, 101-104.

Porter, S. C. (1976) Pleistocene glaciation of the southern part of the North Cascade Range, Washington. *Geol. Soc. Am. Bull.* 87, 61-75.

Porter, S. C. (1981) Pleistocene glaciation in the southern lake district of Chile. *Quaternary Research* 16, 263-292.

Postma, G., Roep, T. B. (1985) Resedimented conglomerates in the bottomsets of Gilbert-type gravel deltas. *J. Sedim. Pet.* 55, 874-885.

Postma, G., Roep, T. B. and Ruegg, G. H. J. (1983) Sandy-gravelly mass flow deposits in an ice marginal lake (Saalian, Leuvenumsche Beek Valley, Veluwe, The Netherlands) with emphasis on plug flow deposits. *Sedim. Geol.* 34, 59-82.

Rappol, M. and Stoltenberg, H. M. P. (1985) Compositional variability of Saalian till in the Netherlands and it origin. *Boreas* 14, 33-50.

Read, K. O. (1963) *The geology of the Princess River area.* BSc(Hons) thesis, Dept. of Geology, University of Tasmania.

Rose, (1974) Small-scale spatial variability of some sedimentary properties of lodgement till and slumped till. *Proc. geol. Ass.* 85, 239-258.

Robertson, N. (1986) *Seismic reflection of shallow bedrock and unconsolidated sediments at King River, western Tasmania.* BSc (Hons) thesis Dept. of Geology, University of Tasmania.

Rust, B. R. (1972) Pebble orientation in fluvial sediments. *J. Sedim. Pet.* 42, 384-388.

Rust, B. R. (1975) Fabric and structure in glaciofluvial gravels. In: *Glaciofluvial and glaciolacustrine sedimentation.* (Ed. by A. V. Jopling and B. C. McDonald) p. 238-248. S.E.P.M. Special Publication No 23.

Rust, B. R. (1978) Depositional models for braided alluvium. In *Fluvial Sedimentology.* (Ed. by A. D. Miall) p. 605-625. Can. Soc. Petrol. Geol. Memoir, 5.

Ruszczynska-Szenajch H. (1982) Depositional processes of Pleistocene lowland end moraines and their possible relation to climatic conditions. *Boreas* 11, 249-260.

Sandford, A. R. (1959) Analytical and experimental study of simple geologic structures. *Geol. Soc. Am. Bull.* 70, 19-52.

Sansom, P. W. (1978) *The glacial geomorphology of the Tullah area.* BSc(Hons) thesis, Dept. of Geography, University of Tasmania. 278 pp.

Selley, R. C. (1970), Studies of sequence in sediments using a simple mathematical device. *Quart. J. Geol. Soc. Lond.* 125, 557-581.

Shackleton, N. J. and Opdyke, N (1973) Oxygen isotope and palaeomagnetic stratigraphy of equatorial Pacific core V28-238: oxygen isotope temperatures and ice volumes on a 10^5 and 10^6 scale. *Quaternary Research* 3, 39-59.

Sharp, M. (1985) Sedimentation and stratigraphy at Ejabakkajokull-An Icelandic surging glacier. *Quaternary Research* 24, 268-284.

Shaw, J. (1971) Mechanism of til deposition related to thermal conditions in a Pleistocene glacier. *J. Glaciology* 10, 363-373.

Shaw, J. (1972) Sedimentation in the ice contact environment with examples from Shropshire, England. *Sedimentology* 18, 23-62.

Shaw, J. (1975) Sedimentary sucessions in Pleistocene ice-marginal lakes. In:*Glaciofluvial and Glaciolacusterine Sedimentation.* (Ed. by A. V. Jopling and B. C. McDonald) p. 281-303. S.E.P.M. Special Publication No 23.

Shaw, J. (1977) Till deposited in arid polar environments. *Can. J. Earth Sci.* 14, 1239-1245.

Shaw, J. (1979) Genesis of the Sveg Tills and Rogen Moraines of central Sweeden: A model of basal melt-out. *Boreas* 8, 409-426.

Shaw, J. (1980) Application of present day glacial processes to the interpretation of ancient tills. *Geografia* 12, 49-55.

Shaw, J. (1981) General discussion. *Annals of Glaciology* 2, 188.

Shaw, J. (1982) Melt-out till in the Edmenton area, Alberta, Canada. *Can. J. Earth Sci.* 19, 1548-1569.

Shaw, J. and Archer, J. (1979) Deglaciation and glaciolacustrine sedimentation conditions, Okanagan Valley, British Columbia, Canada. In *Moraines and Varves. Origin/Genesis/Classification.* (Ed. by Ch. Schluchter.) p. 347-355. A. A. Balkema. Rotterdam.

Shepard, F. P. and Young, R. (1961) Distinguishing between beach and dune sands. *J. Sedim. Pet.* 2, 196-214.

Solomon (1962) The tectonic history of Tasmania. In *The Geology of Tasmania.* J. Geol. Soc. Aust. 9, 310-339.

Sprent, T. C. (1887) Recent explorations of the west coast of Tasmania. *Proc. Roy. Geogr. Soc. Aust.* (Victoria Branch) 51-65.

Stockmarr, J. (1971) Tablets with spores used in absolute pollen analysis. *Pollen Spores* 13, 615-621.

Sturm, M. (1979) Origin and composition of clastic varves. In: *Moraines and Varves. Origin/Genesis/Classification.* (Ed. by Ch. Schluchter.) p. 281-285. A. A. Balkema, Rotterdam.

Suggate, R. P. (1965a) *Late Pleistocene geology of the northern part ofthe South Island, New Zealand.* N.Z. Geol Surv. Bull. 77, 91pp.

Suggate, R. P. (1965b) The definition of interglacial. *J. Geol.* 73, 619-626.

Suggate, R. P. (1974) The Otiran/Aranuian boundary. *N. Z. J. Botany.* 12, 235-237.

Suggate, R. P. (1985a) Where have all the glaciations gone? Proceedings of the second CLIMANZ conference. Publication of the Geology Dept. Victoria University of Wellington. No. 31 p. 47-48.

Suggate, R. P. (1985b) The Glacial/Interglacial sequence of North Westland, New Zealand. *N. Z. Geol. Survey Record* 7.

Suggate, R. P. and Moar, N. T. (1970) Revision of the chronology of the late Otira glacial. *N. Z. J. Geol. Geophys.* 13, 742-746.

Wakefield, N. A. (1975) *Ferns of Victoria and Tasmania.* Field Naturalists Club of Victoria, Griffin Press. 80pp.

Vita-Finzi, C. (1973) *Recent Earth History.* Macmillan, London, 138 pp.

Van de Geer, G. Fitzsimons, S. J. and Colhoun, E. A. (1988 in press) Late Pleistocene vegetation history and palaeoclimate at Newall Creek, western Tasmania. *New Phytologist.*

Waller, G. A. and David, T. W. E. (1904) Report of the Glacial Research Committee. *Report of the Australasian Association of Advancement of Science* 10, 613-617.

Ward, L. R. I. (1909) The Mt. Farrel Mining Field. *Report of the Secretary of Mines for Tasmania for 1908*, 1-120.

Watson, R. A. and Wright, H. E. (1980) The end of the Pleistocene : a general critique of chronostratigraphic classification. *Boreas* 9, 153-163.

Wasson, R. J. (1977) Last Glacial alluvial fan sedimentation in the lower Derwent Valley, Tasmania. *Sedimentology* 24, 781-799.

Weertman, J. (1961) Mechanism for the formation of inner morainesfound near the edge of cold ice caps and ice sheets. *J. Glaciology* 3, 965-978.

West, R. G. (1961) Interglacial and interstadial vegetation in England. *Linnaeus Soc. Lond. Proc.* 172, 437-453.

West, R. G. (1985) Climatic change in the Quaternary - evidence and ideas. *J. Geol. Soc. Lond.* 142, 413-416.

West, R. G. and Donner, J. J. (1956) The glaciations of East Anglia and the east midlands. A differentiation based on stone orientation measurements of the tills. *Quart. J. Geol. Soc. Lond.* 112, 69-91.

Willis, J. H. (1970) *A handbook of plants in Victoria*, vol. 1 (*Ferns, conifers and monocotyledons*), 2nd edn. Melbourne University Press, 481pp.

Young, J. A. T. (1969) Variations in till macrofabric over very short distances. *Geol. Soc. Am. Bull.* 80, 2343-2352.

APPENDIX 1.

Weathering rind data

x= mean
s.d.= standard deviation
Max= maximum weathering rind thickness
Min= minimum weathering rind thickness

Grid references are from Tasmap sheet 8013 Franklin (Fig. 2.1)

Site No.	Field reference	Formation	Grid reference	x	s.d.	Max	Min	Sediment
1	LK 66	Long Marsh	874 309	0.39	0.17	1.1	0.2	Gravel
2	UK		885 422	0.42	0.16	1.1	0.2	Gravel
3	P 63		882 394	1.4	0.6	2.5	0.5	Gravel
4	TP 52		878 307	0.82	0.39	2	0.3	Gravel
5	K 64		881 392	0.5	0.14	0.75	0.2	Gravel
6	Dante	Dante	902 456	1.5	0.22	3	0.8	Till
7	Sedgewick		843 478	1.9	0.2	3	1	Till
8	K 80	Chamouni Valley	899 425	1.53	0.7	2.3	0.8	Till
9	K60		881 396	1.27	0.42	2.2	0.6	Gravel
10	K 62		881 398	1.55	0.42	2.8	0.7	Gravel
11	K 77	Blackwood	896 368	8.68	2.66	16	5	Till
12	K 84		896 369	15.61	3.32	23	9	Till
13	B 1		896 370	13.64	2.9	20	11	Till
14	B 2		896 370	8.84	2.4	13	4	Gravel
15	R 1		875 311	1.95	0.5	2.8	0.5	Gravel
16	G 3		881 306	2.2	0.6	4	1	Gravel
17	TP 48		875 305	2.5	0.64	4.5	1	Gravel
18	VC 1		932 284	9.9	3.1	17	6.5	Till
19	CC 2		871 408	8.03	1.75	12.5	5	Till
20	D1	King	885 349	14.8	9.3	35	5	Till

Site No.	Field reference	Formation	Grid reference	x	s.d.	Max	Min	Sediment
21	K 85		883 315	7.6	1.8	12.5	4	Till
22	TP 60		880 307	5.6	1.1	8	3.5	Gravel
23	TP 58		879 308	4.7	1.1	6.5	3	Gravel
24	TP 54		880 308	9.77	2.8	17	4	Gravel
25	LK 4		882 307	5.3	1.6	8.5	2.5	Gravel
26	LK 5		881 307	5.4	1.9	9.5	2.5	Gravel
27	Bax 1		887 298	5.3	1.4	9	3	Gravel
28	K 54a		875 297	5.1	1.4	8	3	Gravel
29	Rail 1		888 342	6.2	1.8	8.5	4.5	Gravel
30	Bax 2	Governor	878 298	17.3	2.5	20	16	Gravel
31	K 54b		875 297	14.3	2.9	19	8	Gravel
32	G 2	Thureau	878 307	56.6	24.5	96	37	Till
33	K 28		882 351	54.5	28.2	75	39	Till
34	K 27a		882 349	29.8	7.8	54	14	Till
35	K 27b		882 348	33.5	61	61	14	Till
36	N 9		786 318	14.5	6.3	28	6	Gravel
37	R 2		875 311	75.5	14.5	90	35	Till
38	TP 48	Unknown	877 307	2.5	0.63	4	1.5	Gravel
39	Ne 1		917 367	9.4	2.7	13	7	Till

APPENDIX 2.

Quaternary clastic dykes in the King Valley.

A paper to be submitted to Sedimentology.

QUATERNARY CLASTIC DYKES IN THE KING VALLEY, WESTERN TASMANIA, AUSTRALIA.

Sean J. Fitzsimons[1] and Eric A. Colhoun[2].

1. Department of Geography, University of Tasmania, G.P.O. Box 252C, Hobart 7001.

2. Department of Geography, University of Newcastle, Newcastle, N.S.W. 2308.

ABSTRACT.

Quaternary clastic dykes are formed in a variety of sediments in the King Valley. Four types are described and a variety of origins are inferred. They include; till dykes injected into bedrock fractures produced by overriding ice; gravel dykes in bedrock openings that appear to be eroded and stream-filled bedding planes; gravel dykes in weathered limestone that formed as fillings of solution dolines and tunnels and gravel dykes in unconsolidated Quaternary deposits.

The majority of the dykes in the King Valley are gravel-filled, wedge-shaped structures in Quaternary tills and outwash gravels. Most occur in three distinct swarms. The dykes of one swarm have along-slope strikes and are formed on laminated glacial lake sediments that have been subjected to landsliding. The dyke structures appear to have formed as tensions cracks caused by the landslides. The other two dyke swarms have downslope strikes and are associated with sediments deposited in ice-contact environments. It is suggested that the dykes were formed syndepositionally by the collapse and deformation of the sediments as buried ice melted.

INTRODUCTION.

Clastic dykes are vertical or near vertical wedge-shaped structures filled with clastic material which cut through sedimentary rocks (Dionne and Shilts 1974). They have also been termed neptunian dykes, clastic wedges and sedimentary dykes. (Allen 1982) The terminology of this paper follows that of Dionne and Shilts (1974).

The genesis and occurrence of clastic dykes have been reviewed by Hayashi (1966), Dionne and Shilts (1974), and Allen (1982). These reviews show that they are common in rocks of various ages and form in a variety of ways. The numerous suggested modes of origin include; tree root fillings; differential solution of calcareous deposits; fills of tectonic tension cracks; desiccation cracks; thermal contraction cracks including ice wedges; till injection; fills of tension cracks from slumping, hillside creep and landsliding.

Clastic dykes have been observed at several locations in Tasmania (Read 1963, Colhoun 1977, Sansom 1978). They have not been accurately described and their origin is not understood. Between 1982 and 1985, 85 dykes were exposed in excavations in the King Valley. The dykes can be divided into four groups based on the origin of host and fill sediments. Three till dykes in bedrock have formed by injection of basal till into fractures produced by over-riding ice. Four dykes in bedrock appear to be stream fills of opened or preferentially eroded strata. Seven dykes in weathered limestone are filled solution dolines. The remaining 71 dykes are in unconsolidated Quaternary tills and outwash gravels. Four are isolated occurrences and 67 occur in three swarms of 35, 23, and 9 dykes (Fig. 1). The ways in which these dykes were formed is not immediately obvious in the field. Our analysis indicates they have a variety of origins.

The dykes were described, sketched and photographed. Particle size analysis, stone counts and pebble fabric were used to demonstrate the differences and similarities in the host and fill sediments. The strike and dip values of dyke walls were measured to show relationships between

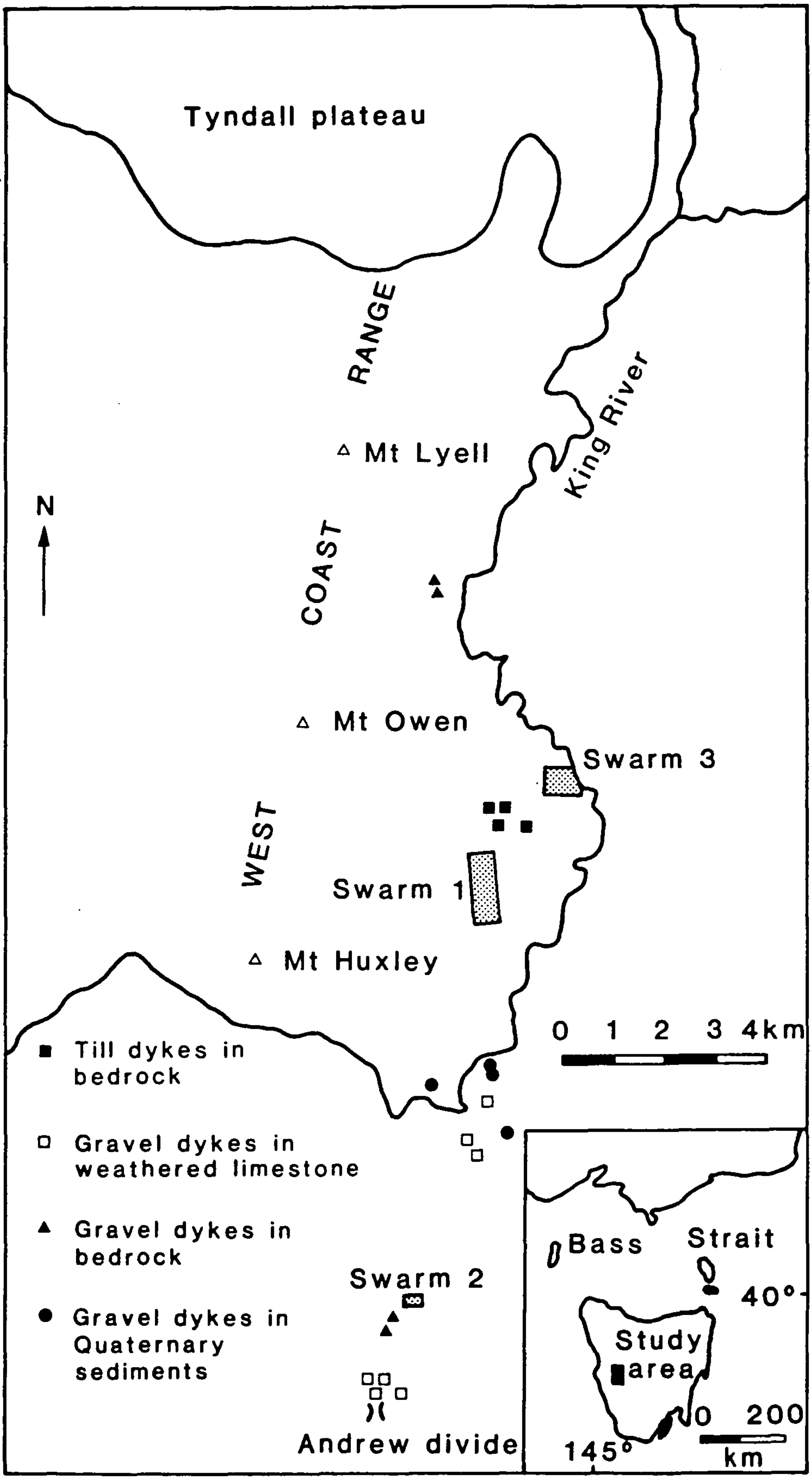


Figure 1. Location map and location of clastic dyke swarms.

dyke structure, slope and host sediment fabric.

This paper outlines the geology of the King Valley, presents and discusses the data obtained during the study of the dykes, and suggests possible origins for their formation.

REGIONAL SETTING.

The King Valley is a structurally controlled valley in a geologically complex area. It lies within a north-south trending belt of faulted Ordovician, Silurian and Devonian rocks (Corbett et. al. 1977). Relatively soft Ordovician and Siluro-Devonian sediments underlie the fault-bounded valley. To the west, resistant Ordovician conglomerate and Cambrian volcanic rocks form the West Coast Range (Fig. 1), and to the east faults separate Devonian sediments from Precambrian metasediments A plateau composed of Precambrian metasediments, Permian sediments and Jurassic dolerite forms the headwaters of the valley.

During the Quaternary the valley was glaciated several times (Colhoun 1985a). Ice from the Tyndall plateau flowed south and formed an outlet glacier in the King Valley. At its maximum extent the glacier terminated at the entrance to the King River gorge through the West Coast Range (Fig. 1).

Unconsolidated Quaternary deposits cover most of the valley floor. The ages of the glaciations are not yet certain but deposits of the oldest or Linda Glaciation are thought to be of early Pleistocene age because lake laminated sediments have a reversed detrital remanent magnetisation (M. Pollington pers. comm. 1987). Most of the glacial deposits in the King valley are of Comstock Glacial age. They are probably of middle Pleistocene age because they are beyond the range of radiocarbon dating, underlie an interglacial deposit, are normally magnetised and are much less chemically weathered than the Linda age deposits. Last or Margaret Glaciation age deposits are confined to the high cirques of the West Coast Range and the Tyndall Plateau at the northern end of the valley.

Although there is geomorphological and biogeographical evidence that the region has experienced several climatic changes during the Quaternary (Kiernan 1983; Colhoun 1985a, 1985b; Colhoun and van de Geer 1986) the data suggest that maximum temperature depression was only about 6-7° C below present mean annual temperature. The region had a moist climate throughout the Quaternary, even though it may have been drier than present. There is no evidence from western Tasmania either for extreme cold such as would be consistent with the development of permafrost conditions, or for subarid or arid conditions. Pollen data indicate cool, moist, nearly treeless herbland, heathland and shrubland environments during the colder periods and rainforest during the warmer periods.

RESULTS

The characteristics of the four types of dyke are summarised in Table 1. Till dykes penetrating Devonian quartzite are exposed in a quarry on the upstream end of a roche moutonnee (Fig. 2A). The narrow, steeply-dipping dykes are all irregular in outline and three become wider with depth. Rudimentary flow structures indicate that the dykes were filled from above. As the dykes were not fully exposed the depth of bedrock penetration is unknown, but was at least 2.5 m.

Gravel dykes in Devonian quartzite and Ordovician conglomerate are shallow wedge-shaped structures parallel to the vertically dipping bedding planes of the host rock (Fig. 2B). Fills of these dykes in the southern part of the valley consist of rounded, fluvial pebble gravel with rare pieces of angular host rock, while those in the north are filled with angular colluvial gravels (Fig. 1). The extent of the dykes is unknown but one has been traced for 25 m.

Gravel dykes in black clayey silt, the weathering product of Ordovician limestone (Banks 1957; Read 1963), have three forms. Dykes up to 20 m wide with gently dipping walls are filled solution dolines. Small dykes up to 4 m wide with steep nearly vertical walls (Fig. 2C) are solution pipes of former dolines. Shallow wedge-shaped structures with clasts dispersed in a

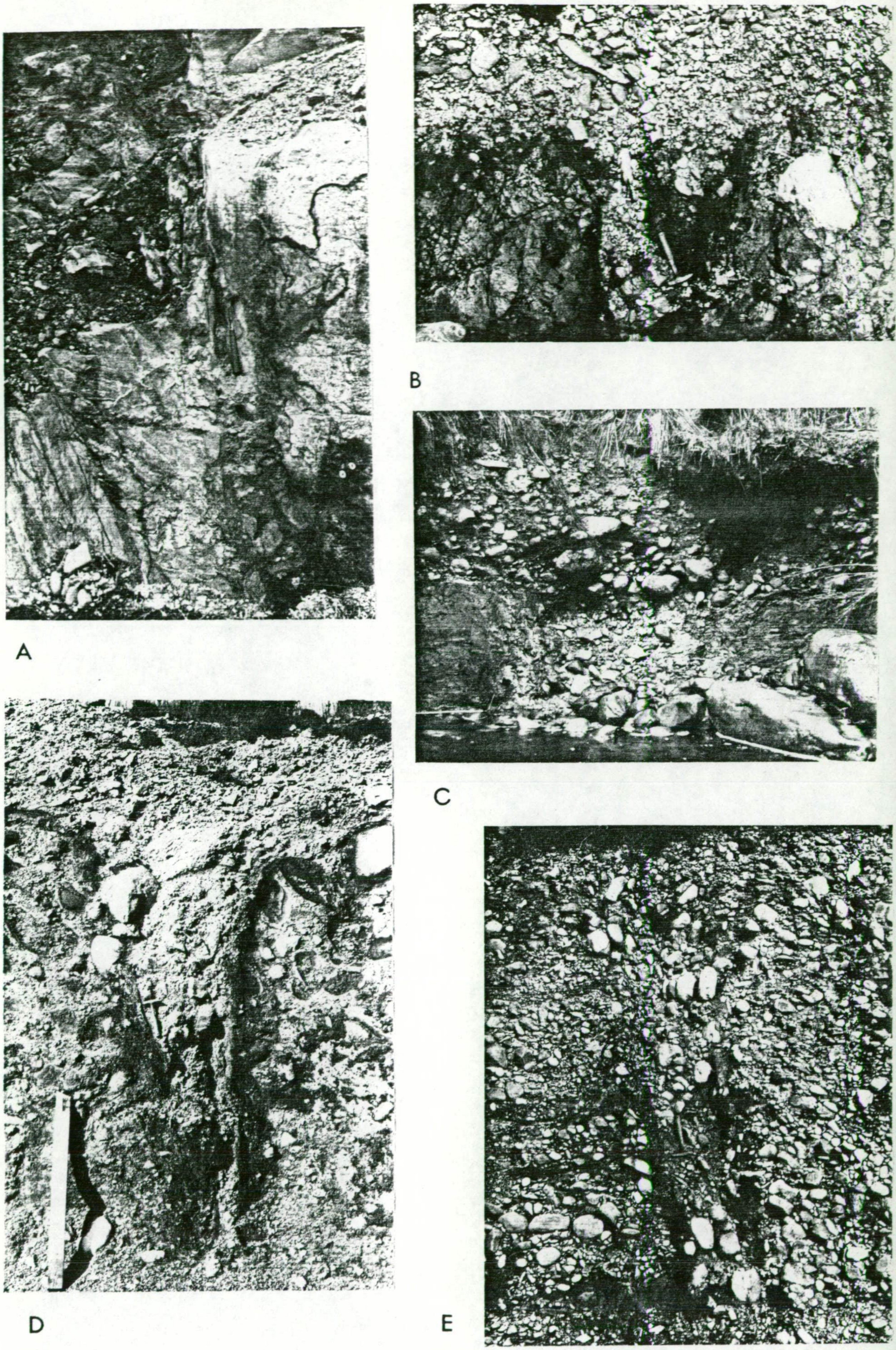


Figure 2. (A) 2.5 m till dyke in Devonian quartzite. (B) 1 m Gravel dyke in Devonian quartzite. (C) 1.1 m gravel dyke in Devonian quartzite. (D) 3.6 m dyke in weathered till. Note the vertical layering in the marginal black stained sand. (E) 4 m gravel dyke in ice contact gravels. Note the arcuate fabric at the top and sharp truncation of strata at the bottom.

TABLE 1 Summary of dyke characteristics.

Dyke type	Dimensions	Dip and strike patterns	Structures	Host sediment	Fill sediment	Mode of origin
Till dykes	Up to 150 cm wide and 2.5m deep	Steep dips (40-80^{o}). Strike is oblique to striations	Rudimentary flow structures related to fluid injection	Well indurated Devonian quartzite	Fine grained till	Injection of basal till into pre-existing glaciotectonic fractures.
Gravel dykes in noncalcareous bedrock	Up to 100 cm wide and 1.5m deep.	Near vertical dip Strike is transverse to bedding planes and pebble fabric of host sediment.	Fills massive and host rock brecciated.	Poorly indurated Devonian quartzite and well indurated Ordovician conglomerate	Outwash gravel and slope deposits	Erosion and stream fills of bedding planes.
Gravel dykes in weathered limestone	Large size range Up to 5m wide and 6m deep	Sides are generally steep.	Vertical clast alignment at the dyke margins	Weathered limestone calcareous siltstone	Outwash gravel and till	Filled solution dolines and tunnels
Gravel dykes in Quaternary tills and outwash gravel	Up to 4m deep and 1.5m wide	Near vertical. Strike is oblique to pebble fabric of host sediment, parallel and transverse to slope.	Vertical clast alignment. Marginal sands laminated suggesting a multi phase fill	Melt-out and flow tills. Ice contact stratified gravel.	Alluvial fan and slope deposits	Several possible origins including:- 1 Mass movement 2 Syndepositional ice melt deformation 3 Ice wedges 4 Glaciotectonic tension crack 5 Tectonic tension crack

matrix of weathered limestone appear to be collapsed solution tunnels. Numerous drill logs show that gravel lenses and open tunnels occur in the weathered limestone up to 14 m below ground level in the area of the Andrew Divide. There is little doubt that these dykes represent a filled karstic drainage system.

Dykes also occur in early Pleistocene tills (swarm 1, Fig 1 & Fig 2D), in middle Pleistocene tills and outwash gravels (swarm 3, Fig 1), and in ice contact-contact stratified gravel of uncertain Pleistocene age (swarm 2, Fig. 1 and Fig. 2E). The structure and composition of these wedge-shaped dykes are summarised in Table 1.

Differences between particle sizes of host and fill sediments in swarm 1 (Fig. 3A) reflect different modes of origin, provenance and age. The host is a highly weathered fine grained till derived from a mixture of rock types, Figure 4B, and the fill is a fan deposit of locally derived gravel from the siliceous rocks, Fig. 4A. Particle sizes and lithology of host and fill sediments in swarms 2 and 3 are almost identical because fills are derived from the host sediment (Fig. 3B, 3C and Fig. 4C-F). Dyke fills of swarms 2 and 3 are slightly finer than the host because they contain recently eluviated mud.

Layers of coarse and medium sand occur at the margins of several dykes (Figs. 2D and 2C). The layers are 5 to 15 mm. thick, have sharp boundaries and a planar form. Although their origin and significance is unclear, they may be the result of several episodes of filling as the dykes opened repetitively. Similar "phase" and "pulse" layering of biogenic debris in limestone dykes has been interpreted as different periods of filling by Lewis (1973).

Lower hemisphere equal area projection diagrams of pebble fabric summarise the direction of transport of host and fill sediments (Fig. 5). They are contoured using Kamb's (1959) method with an interval of 2o and a significance level of 3o. The direction of maximum clustering (V_1) was calculated by the eigenvalue method discussed by Mark (1973). Only prolate-shaped pebbles with approximately equal b and c axes and an a/b ratio of 2 or greater were used. Figure 5 shows

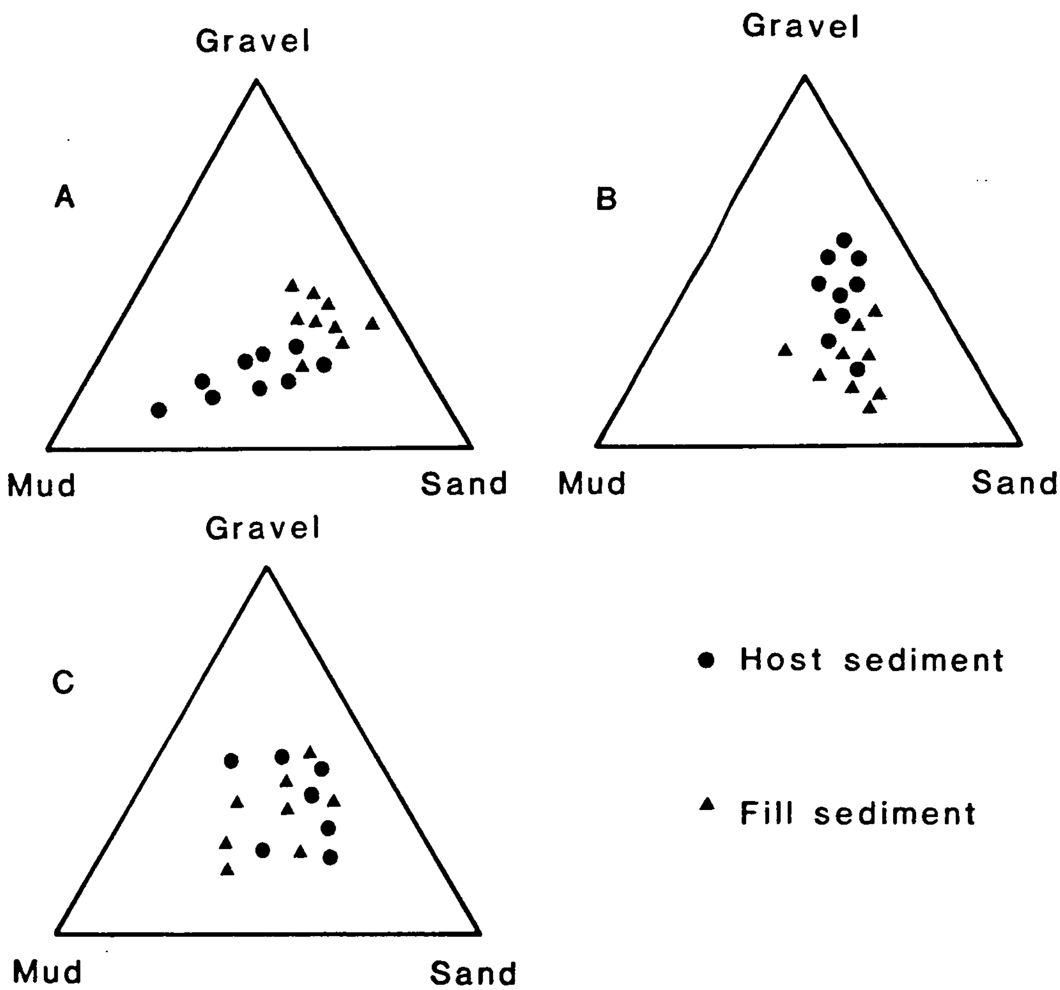


Figure 3. Ternary diagrams of dyke host and fill particle sizes. (A) Swarm 1. (B) Swarm 2. (C) Swarm 3.

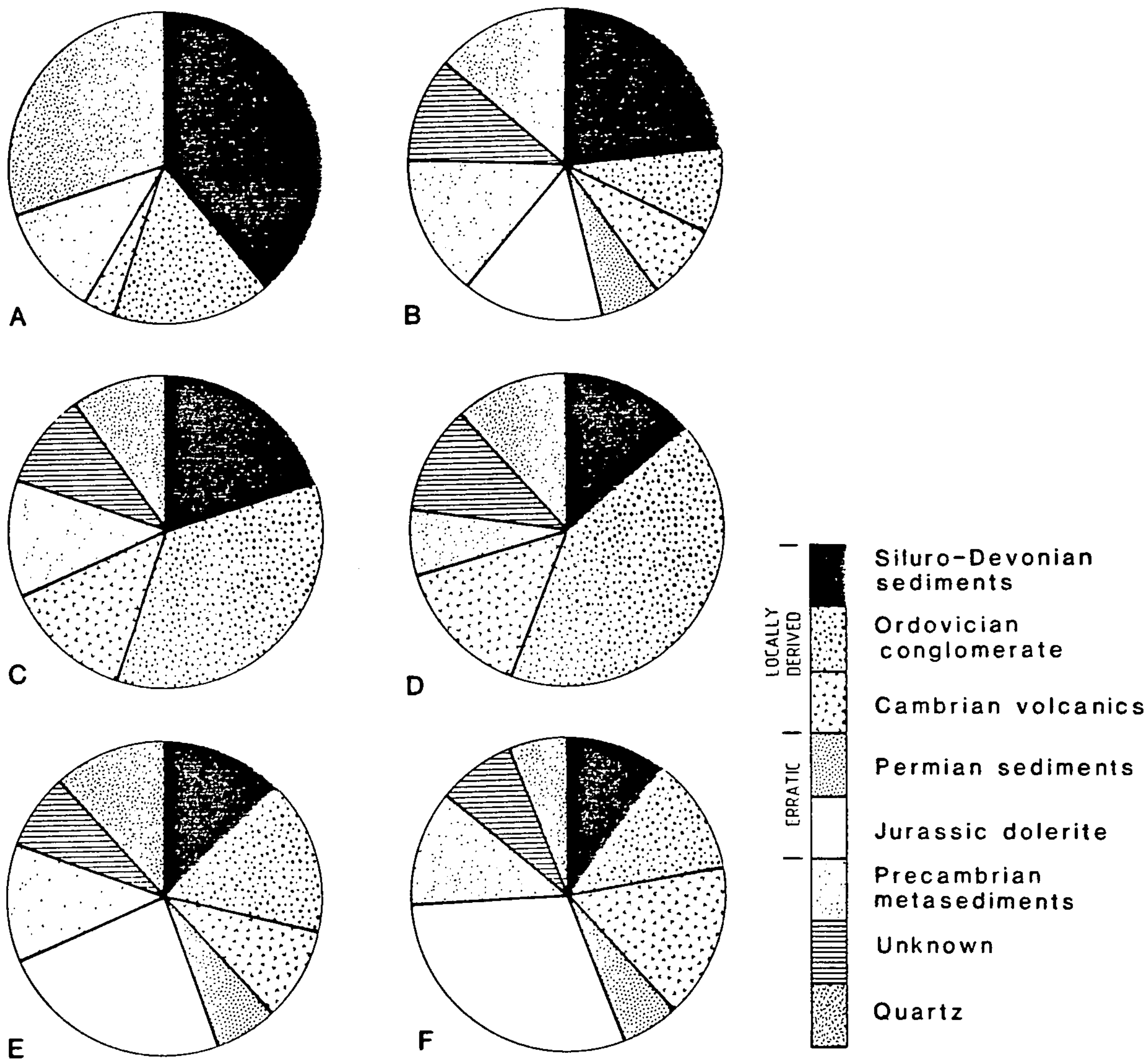


Figure 4. Lithology of dyke host and fill sediments. (A) Fill of swarm 1. (B) Host of swarm 1. (C) Fill of swarm 2. (D) Host of swarm 2 (E) Fill of swarm 3. (F) Host of swarm 3.

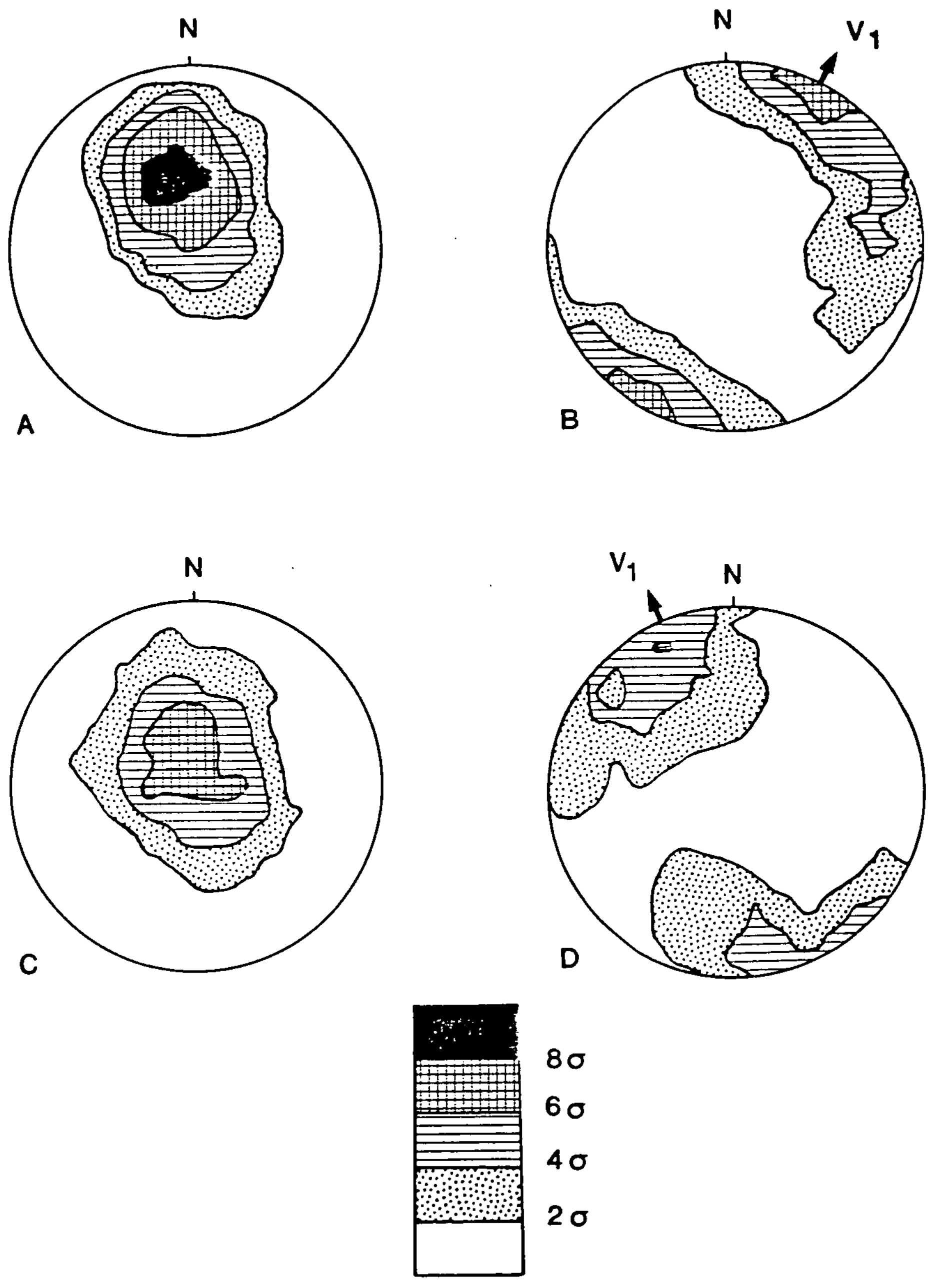


Figure 5. Contoured equal area projections of pebble fabric of dyke host and fill sediments. (A) Fill of swarm 1. (B) Host of swarm 1. (C) Fill of swarm 2. (D) Host of swarm 2.

that fill sediments have strong, steeply-dipping fabric concentrations developed by the filling from above.

Figure 6 shows the relationship between dyke strike, slope and maximum clustering of host sediment fabrics. Dykes from swarms 1 and 2 are parallel to the surface slope and dykes from swarm 3 strike across the slope.

All clastic dykes in the unconsolidated Quaternary deposits are associated with zones of humus and iron staining along the dyke walls (Fig. 2D and 2E). The staining frequently forms unfilled cracks which may be an early stage in dyke development or a poorly developed dyke. Absence of pollen from the humus-iron-enriched sediment suggests that it was not open to the atmosphere during formation. It appears to have been formed by the leaching of humic acids from a surface soil and their concentration in the wedge structure adjacent to the largely impermeable host materials. The present pedogenic regime is clearly one of strong podsolisation which would be capable of mobilising and precipitating the humus-iron colloids.

DISCUSSION.

Two, only partly separable, considerations in the formation of clastic dykes are the initiation and propagation of cracks and their filling. For some types crack formation and filling occur simultaneously (Worsley 1975; Harms 1965), for others a substantial period of time separates crack formation and filling. Allen (1982) considers that many sedimentary intrusions arise from liquefaction of sediments, but clastic dykes may also form when overlying sediment falls into an open or opening tension crack. The modes of origin of cracks and their fills are considered for the four types of dykes with emphasis being given to the dykes in the unconsolidated deposits.

Till Dykes in Bedrock.

the occurrence of dykes in high stress areas on the upstream slope of a roche moutonnee together

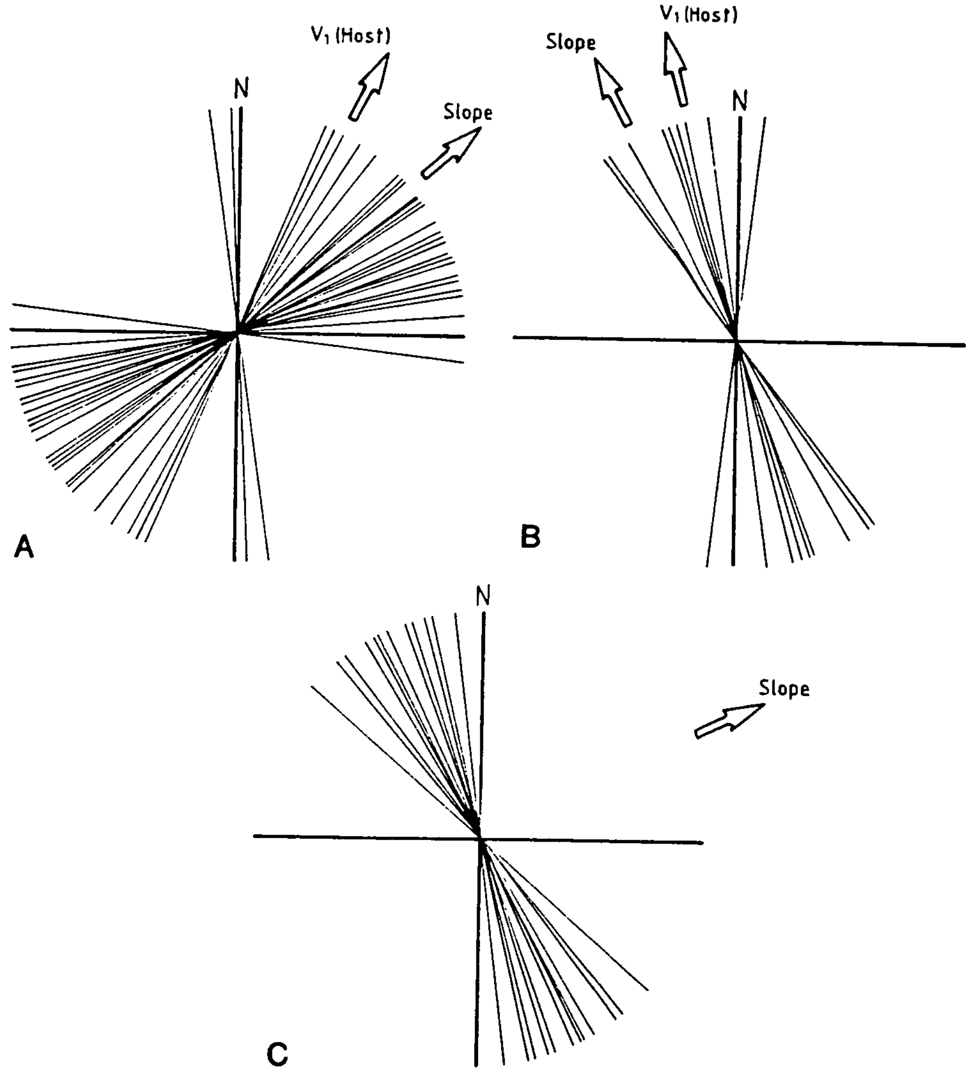


Figure 6. Strike of dykes from swarms 1, 2 and 3 in relation to surface and palaeoslopes and host sediment fabric. (A) Swarm 1. (B) Swarm 2. (C) Swarm 3.

with flow structures in the fill indicate that their emplacement was by injection of saturated basal till from an over-riding glacier.

Although till commonly fills irregularities in bedrock, there are few records of till injected into consolidated rocks. Black (1983b) described an example of till dykes penetrating a granite roche moutonnee in Connecticut.

There are two differences between the dykes described by Black and those which occur in the King Valley. Dykes in the King Valley are up to 3 times wider and become wider with depth while those described from Connecticut become narrower with depth. The King Valley dykes occur in strongly jointed rocks at a site which is estimated to have been covered with at least 400 m of ice. The dykes appear to be located where large blocks of rock have been fractured and rotated by over-riding ice, thereby producing cracks of irregular size and shape. Although the means of crack initiation is unknown, it is likely that the rock responded to glacial unloading in a manner similar to that suggested for Connecticut by Black (1983b). Hence, the joints were probably opened by dilation during deglaciation and the fill of saturated basal till either flowed or was injected into the open crack.

Gravel Dykes in Bedrock.

Because gravel dykes in bedrock have been formed parallel to the steeply dipping bedding planes of the host rock, it appears that formation of the initial crack was related to rock structure. Although the mechanism of crack formation is unclear, it appears that crack formation and fill were contemporaneous because the fill incorporates stream gravels mixed with angular fragments of host rock.

There are few descriptions of gravel dykes in consolidated rocks. Hayashi (1966) describes conglomerate dykes in sandstone but does not comment on their origin other than to say they are filled rather than injected or intruded. Stewart (1911) describes conglomerate dykes in Arizona similar to the gravel dykes in the King Valley. The dykes in Arizona occupy joints enlarged by weathering, and are filled partly with the product of weathering and with stream sediments. The dykes of the King Valley have a similar mixed fill that occurs parallel to the bedding rather than in joints. The mechanism of crack formation is unclear. Either weathering and erosion, or dilation following the removal of glacier ice may have formed them even though the fills have resulted from fluvial deposition. The fluvially transported fills suggest that weathering and erosion along bedding planes is the more likely hypothesis.

Gravel dykes in Weathered Limestone

Gravel dykes in weathered limestone are filled underground channels and solution dolines similar to those described as streamsink dolines by Jennings (1985). The gravel fillings are either part of a former karst system or glaciofluvial deposits that have plugged a pre-existing karstic system.

In the King Valley karstic limestone is overlain by thick deposits of black clayey silts which, although they may be a weathering product of limestone, may have developed from a transitional calcareous siltstone that overlay the Ordovician limestone (F. J. Baynes pers. comm. 1985). It is in this clayey silt that most of the gravel dykes are formed.

The gravel fills are structureless except close to the dyke walls where vertical clast orientation formed by filling from above is preserved.

Almost all these dykes are filled with gravel from a middle Pleistocene advance of the Jukes Glacier, which buries an extensive karstic drainage system. Lack of surface karst features in the area today suggests that the channels of the old karst drainage system have not been used since they were filled in middle Pleistocene times.

Gravel Dykes in Unconsolidated Quaternary Sediments.

Dykes from swarm 3 occur in the upper 5 m of an 80 m-thick sequence of middle Pleistocene sediments that include 44 m of laminated mud. Deformation structures in the mud, and a series of ridges, hollows and back-tilted blocks on the surface of the deposits that slope towards the King River, suggest that successive landslides have occurred. The dykes strike across this slope (Fig. 6c), and appear to be filled tension cracks that have developed in the upper slopes of successive landslides.

There are many recorded instances of mass movements generating tension cracks and sedimentary dykes (Allen 1982). Black (1983a) described clastic dykes in a slope parallel and *en echelon* array that formed as fills of tension cracks produced by intermittent soil creep. The dykes of swarm 3 appear to be similar to those described by other authors and are clearly related to mass movements, probably mainly of the rotational slump type.

The origin of dykes from swarms 1 and 2 are difficult to explain. Some modes of origin that have been suggested for the formation of clastic dykes are clearly not applicable. The dykes are too wide and too continuous to be tree root filings. Linear shrinkage of the gravels (1-4%) and tills is too low, and the dykes are too wide to be desiccation cracks. As these dykes do not overlie limestone they cannot be due to solution subsidence.

Other possible modes of origin that cannot be ruled out easily are; syndepositional ice melt, ice wedge growth, glaciotectonic tension cracks and tectonic tension cracks.

A variety of syndepositional injection, diapiric, and collapse structures occur in sediments deposited in and on ice (Shaw 1972). Dykes from swarm 1 and 2 occur in sections that have numerous syndepositional ice melt features (Figs. 7 and 8). A section in swarm 1, shows numerous high angle reverse faults and a large wedge shaped structure that has been downfaulted by 1.3 m (Fig. 7). On the southern side of the section the silts are highly deformed and

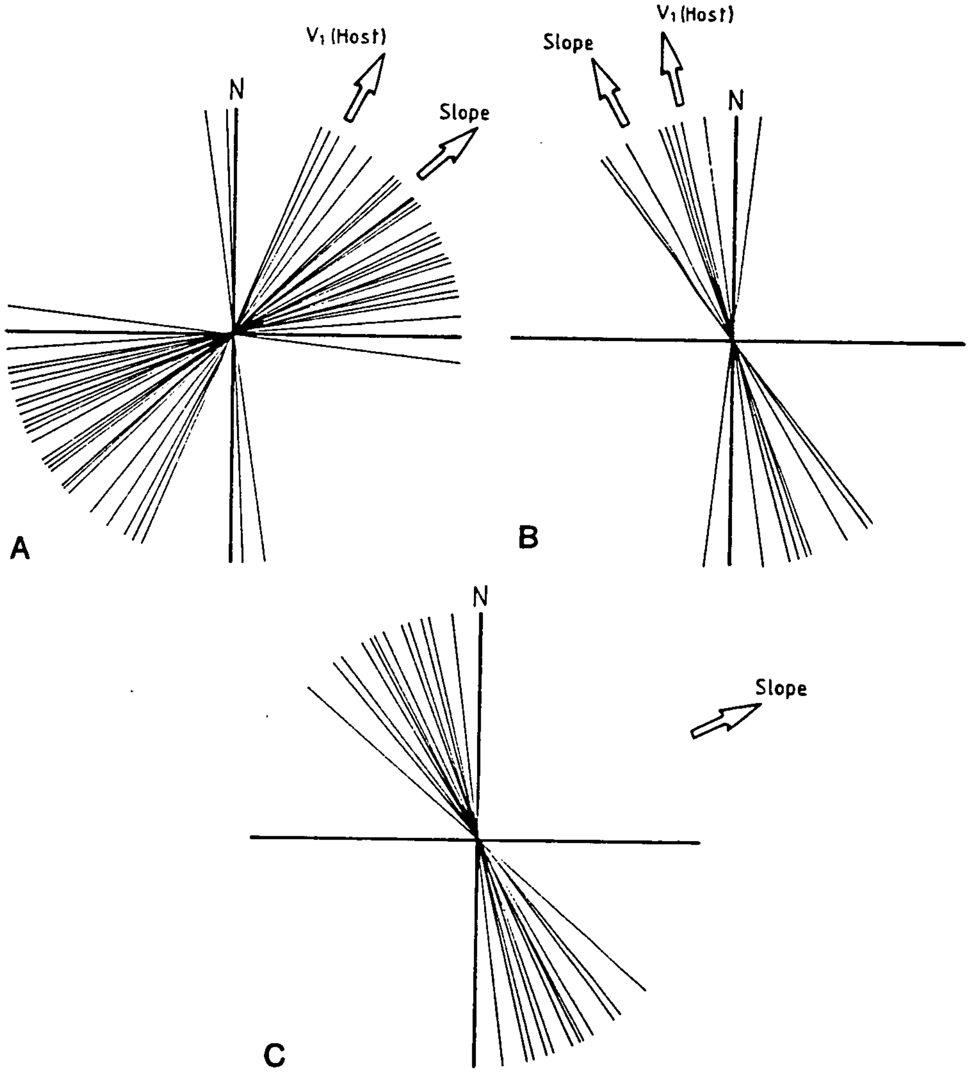


Figure 6. Strike of dykes from swarms 1, 2 and 3 in relation to surface and palaeoslopes and host sediment fabric. (A) Swarm 1. (B) Swarm 2. (C) Swarm 3.

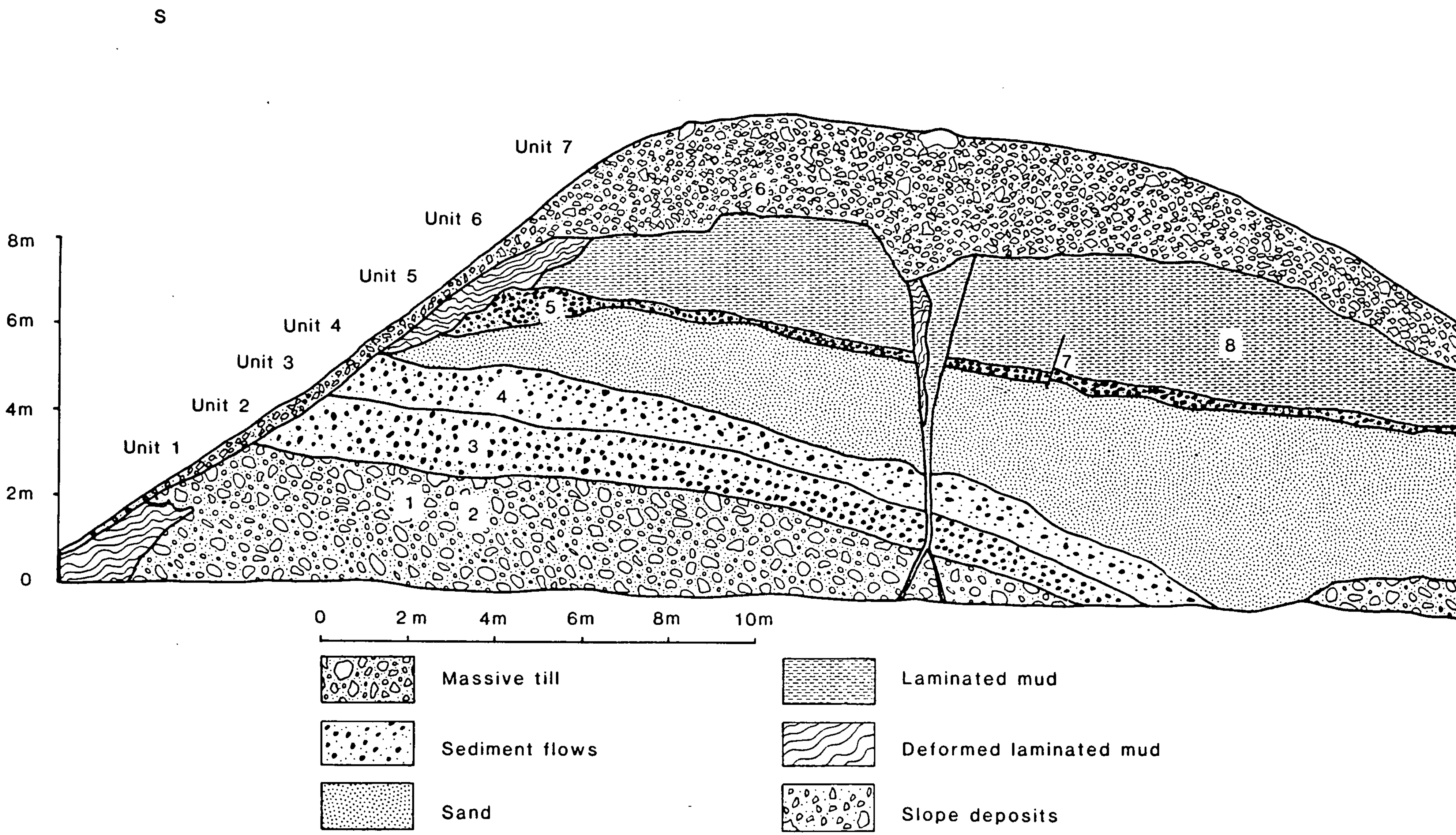


Figure 7. Section through deformed supraglacial sediments in swarm 1. Sediments consist of multiple flow tills, massive sand, flow till, laminated silts and slope deposits, all of which are deformed by melt of buried ice.

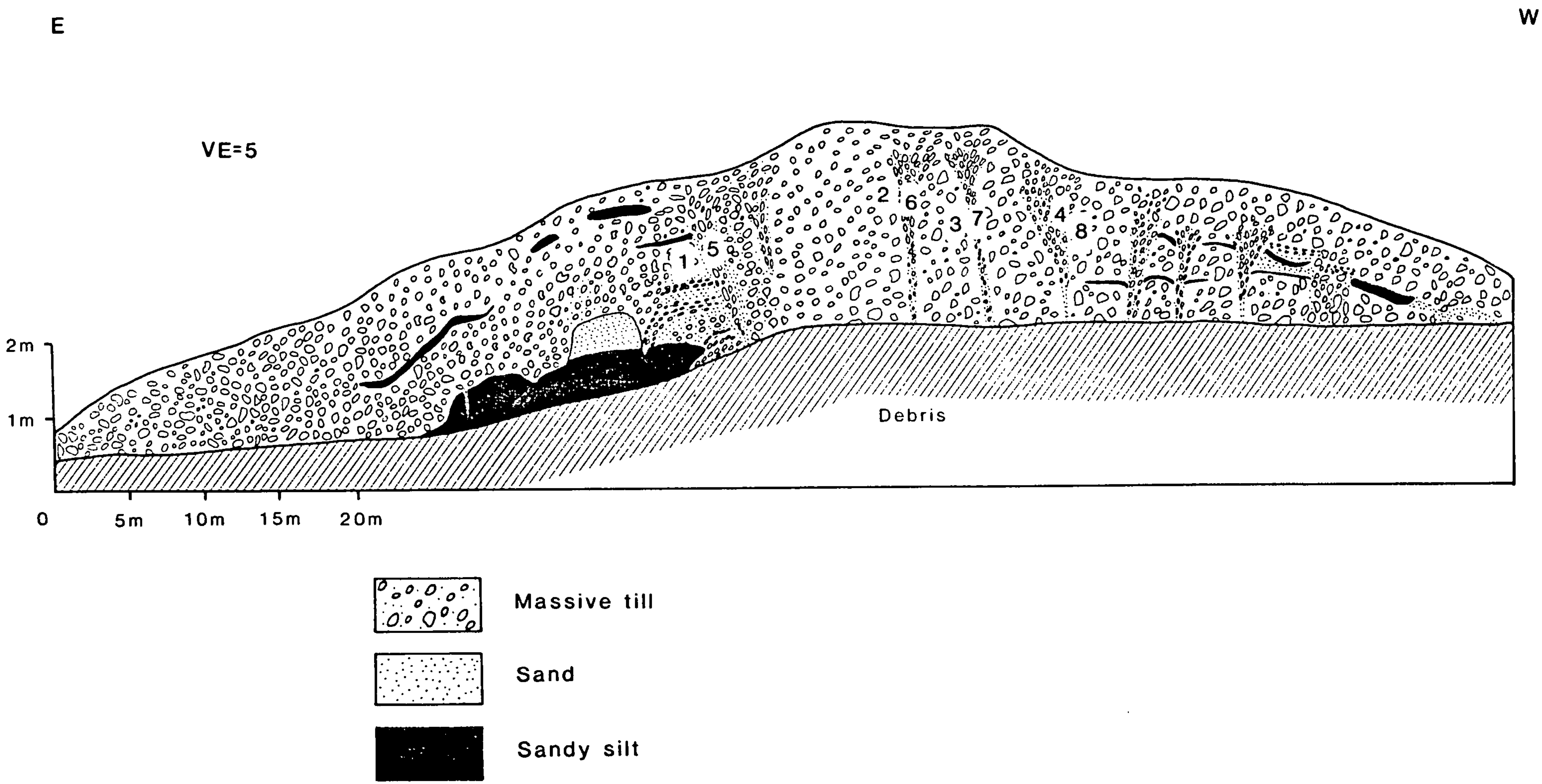


Figure 8. Section through ice contact outwash gravels of swarm 3.

downwarped by 5 m. The deformation has probably been caused by the collapse of sediments into voids created by melting ice blocks and by the slumping of coherent masses of sediment on an an ice surface. Deformation of this type is described by Sandford's (1959) type 1 and 2 experiments which have been used in the interpretation of faults in glaciofluvial sediments by McDonald and Shilts (1975). The models explain both the numerous parasitic, high-angle, reverse faults and the decreased dip of the reverse faults toward the surface.

The dykes of swarm 2 are also associated with zones of collapse and sedimentary structures that suggest deposition on melting ice (Fig. 8).

Clastic dykes may form in a variety of ways in ice contact environments. These include narrow zones of collapse that extend upwards from blocks of melting ice and tension cracks from slumping of sediment at the till/ice interface. Because the dykes occur in deformed supraglacial sediments and strike across the palaeoslope of sediment flow surfaces (Fig. 6A), they may have formed as tension cracks due to mass movement. Similar laterally extensive, boulder filled wedges have been observed to form in supraglacial debris on Icelandic glaciers by a process of lateral migration of thaw lakes that causes destabilisation and mass movement of till slopes (Eyles 1979). If the dykes formed in this way, the slope deposits that overlie the glacigenic sediments of swarm 1 must have buried supraglacial sediments on the margin of the melting King Glacier. This agrees with other observations that have suggested the slope deposits were of paraglacial origin, that is, their formation was conditioned by glaciation. They were deposited throughout the central West Coast Range during and immediately after deglaciation, and were sometimes deposited on melting glacier surfaces.

Several scientists have commented on the physical similarities between the clastic dykes of swarm 1 and ice-wedge casts in Europe (F. J. Baynes and R. G. West pers. comm. 1985). Ice wedge casts are diagnostic of former permafrost but are difficult to identify with certainty in areas without independent evidence of former permafrost. Misidentification and controversial interpretations are numerous (Black 1976, 1983a). Table 2 compares the characteristics of ice

TABLE 2 Comparison of the characteristics of ice wedge casts and clastic dykes in the King Valley.

Ice wedge characteristics	Characteristics of the King Valley dykes in unconsolidated deposits
Wedge shaped	Wedge shaped
Occur in polygonal arrays (Black 1976)	No polygonal patterns. Consistent parallel strike within swarms.
Pressure effects from ice expansion including vertical clast alignment and upturned strata. (Black 1976)	No pressure effects. Strata are sharply truncated and slumped downward.
Narrower in coarser sediments. (Black 1983a)	Fine grained tills are host to narrower dykes than occur in coarser gravels.
Smaller secondary and tertiary wedges and polygons. (Black 1974)	Wide range of sizes, no evidence of systematic variations related to different stagers of growth.
Arcuate downward slump structures in fills. (Black 1976)	All fills have arcuate downward slump structures.
Filled from above and preserve vertical clast alignment. (Johnsson 1959)	All dykes are filled from above and have strong vertical clast alignment.
Crack spacings 2 to 3 times crack height. (Lachenbruch 1962)	Dyke spacing highly variable,approximately equal to crack height in swarms 2 and 4, and up to 10 times dyke in swarm 1.
Permafrost necessary for initiation and growth, and associated with cryoturabation. (Black 1976)	No evidence of permafrost or cryoturbation.
Mean annual temperature of below -3.5°C necessary for initiation. (Hamilton et al. 1983)	Climate at the time of formation unknown.

wedge casts and the clastic dykes of the King Valley.

Structures that have been interpreted as ice wedge casts have been described from latitude 41°S in Argentina by Galloway (1985). Although the King Valley occurs at 42°S there is no other geomorphological or biogeographical evidence to suggest a palaeoclimate that would have supported lowland permafrost conditions in western Tasmania. To suggest the occurrence of mean annual temperatures of below -3.5°C. that would be necessary to initiate thermal contraction cracking (Black 1976; Hamilton, Ager & Robinson 1983; Lachenbruch 1962) seems much too extreme.

Although some characteristics of the King Valley clastic dykes are similar to ice wedge casts, the absence of supporting evidence for permafrost or polygonal patterning suggests the structures are not ice wedge casts.

Tension cracks and clastic dykes are frequently generated by tectonic events. Two types of glaciotectonic stress that may have produced tension cracks in the King Valley are; glacioisostatic rebound following deglaciation, and dilation of rocks from glacial loading and unloading. It is not possible to separate these effects by examination of surficial sediments. The effect of glacioisostasy in producing tension cracks is hypothetical. We know of no record of tension cracks being attributed to glacioisostatic unloading.

Local unloading of pressure on rocks beneath glacier ice was documented by Lewis (1954) and invoked to explain fracturing of rock prior to till injection on a roche moutonnee (Black 1983a). Eyles and Paul (1983) describe "pop up" structures in Ontario and New York where in situ decoupling and arching of the upper 3 m. of limestone occurs in response to unloading of glacier ice. They state that such phenomena are seldom seen in glaciated terrain because of the thick veneer of unconsolidated sediment overlying the rockhead. Although the effect of such "pop up" structures on overlying unconsolidated sediments has not been described, it may be similar to the vertical step uplift mechanism of clastic dyke formation described by Harms (1965).

The geophysical effects of a minimum 300 m-thick ice load on the basement rocks of the King Valley are unknown. Although a glaciotectonic origin for the King Valley dykes is possible there is no specific evidence which supports such an interpretation.

Tectonic events, including earthquakes are known to produce tension cracks in unconsolidated sediments (Allen 1982). However, most involve sedimentary intrusion by liquefaction and injection from sources below the cracks (Reimnitz and Marshall 1965; Peterson 1968; Thorson et.al. 1986). Harms (1965) described a model of clastic dyke formation based on vertical step uplift causing progressive extension of reverse faults, opening of a tension crack and filling from above. Although the model of Harms (1965) is a possible mechanism of clastic dyke formation, there are no sedimentary structures that indicate a seismic origin for the King Valley dykes.

Consideration of the form of dyke structures associated with the various possible mechanisms discussed suggests that the dykes of swarms 2 and 3 were probably formed by collapse and flowage of sediments during the melting of ice immediately after deglaciation of the locality.

CONCLUSION.

Quaternary dykes in the King Valley form in a variety of ways. The following modes of origin seem to have been clearly established for the following dyke types in the King Valley:

1. Till dykes in bedrock formed by injection of saturated basal till into fractures produced by over-riding ice.
2. Gravel dykes in non-calcareous bedrock are stream fills of eroded bedding planes.
3. Gravel dykes in weathered limestone are filled solution dolines and tunnels.
4. Gravel dykes of swarm 3 are filled tension cracks generated by landsliding.

The origins of dykes in swarms 1 and 2 have not been clearly determined. They were certainly not formed by the infilling of tree roots, are not related to shrinkages of host materials and are not formed by the solution of limestone. Consideration of other possible origins in conjunction with the form and pattern of the wedge structures strongly suggests they were not formed as ice wedge casts, as structures due to glacioisostatic or rock surface unloading or other crustal mechanisms. The most likely explanation seems to be that most of the wedge structures (58) of swarms 1 and 2 were formed in association with the collapse and flowage of saturated supraglacial sediments in an ice-contact depositional environment where thick sediments mantled buried ice masses that melted gradually away. There is no diagnostic evidence to indicate a specific origin. However, there is strong circumstantial evidence which seems to eliminate their interpretation as ice wedge casts and suggests that their origin is probably related to syndepositional deformation associated with deposition on a melting ice surface. Although there is no specific evidence to support a tectonic tension crack origin, such a possibility cannot be ruled out.

ACKNOWLEDGEMENTS.

Discussions with Dr. Max Banks, Geology Department University of Tasmania; Dr. Fred Baynes, Hydro Electric Commission of Tasmania; and Hazel and Professor Richard West, Botany School Cambridge University are much appreciated. The support of this work by the Australian Research Grants Committee and the Hydro Electric Commission of Tasmania are gratefully acknowledged.

REFERENCES.

Allen, J.R.L. (1982) *Sedimentary structures. Their character and physical basis.* Vol. 1. 663 p. 23 Amsterdam.

Banks, M. R. (1957) *The stratigraphy of Tasmanian limestones.* In:: Limestones in Tasmania. (Ed. by T. D. Hughes.) Geol. Surv. Min. Res. No. 10 Department of Mines, Tasmania.

Black, R. F. (1974) Ice wedge polygons of northern Alaska. In: *Glacial Geomorphology.* (Ed. by D.R. Coates) pp.247-275. New York University Press, New York.

Black, R. F. (1976) Periglacial features indicative of permafrost: ice and soil wedges. *Quaternary Research* 6, 3-26.

Black, R.F. (1983a) Pseudo ice-wedge casts of Connecticut, N.E. United States. *Quaternary Research.* 20, 74-89.

Black, R.F. (1983b) Clastic dykes of basal till, southern Connecticut, attributed to thick stagnant ice. *Geol. Soc. Am. Abstr. Prog.* 15, 125.

Colhoun, E. A. (1977) Late Quaternary fan gravels and slope deposits at Rocky Cape, northwestern Tasmania: their palaeoenvironmental significance. *Pap. Proc. Roy. Soc. Tas.* 111, 13-27.

Colhoun, E. A. (1985a) Glaciations of the West Coast Range, Tasmania. *Quaternary Research.* 24, 39-59.

Colhoun, E. A. (1985b) Pre-last Glaciation maximum vegetation history at Henty Bridge, western Tasmania. *New Phytologist,* 100, 690-699.

Colhoun, E. A., & van de Geer, G. (1986) Holocene to Middle Last Glaciation vegetation history at Tullabardine Dam, western Tasmania. *Proc. R. Soc. Lond.*, B 229, 177-207.

Corbett, K. D. , Green G.R. & Williams, P. R. (1976) The geology of central western Tasmania. In: *Landscape and Man.* (Ed. by M. R. Banks and J. B. Kirkpatrick) Roy. Soc Tas.

Dionne, J. C. & Shilts, W. W. (1974) A Pleistocene clastic dyke , upper Chaudiere Valley, Quebec. *Can. J. Earth Sci.* 11, 1594-1605.

Eyles, N. (1979) Facies of supraglacial sedimentation on Icelandic and alpine temperate glaciers. *Can. J. Earth Sci.* 16, 1341-1361.

Eyles, N., & Paul, M. A. (1983) Landforms and sediments resulting from former periglacial climates. In: *Glacial geology. An introduction for engineers and earth scientists*. (Ed. by N. Eyles) Pergamon Press Oxford.

Galloway, R. W. (1985) Fossil ice wedges in Patagonia and their palaeoclimatic significance. *Zeitschrift fur Geomorpholgie* N.F. 29, 389-396.

Hamilton, T. D., Ager, T. A., & Robinson, S. W. (1983) Late Holocene ice wedges near Fairbanks, Alaska, U.S.A. : Environmental setting and history of growth. *Arctic Alpine Res.* 15, 157-168.

Harms, J. C. (1965) Sandstone dykes in relation to faults and stress distribution in the southeastern Front Range, Colorado. *Geol. Soc. Am. Bull.*, 76, 981-1002.

Hayashi,T. (1966) Clastic dikes in Japan. *Jap. J. Geol. Geogr.* 37, 1-20.

Jennings, J. N. (1985) *Karst geomorphology*. Blackwell. 293 pp.

Johnsson,G. (1959) True and false ice-wedges in southern Sweden. *Geogr. Annaler.* 41, 15-33.

Kamb, W. B. (1959) Ice petrofabric observations from Blue Glacier, Washington, in relation to theory and experiment. *J. Geophys. Res.* 64, 1891-1909.

Kiernan, K. W. (1983) Weathering evidence for an additional glacial stage in Tasmania. *Austr. Geogr. Stud.* 21, 197-220.

Lachenbruch, A. H. (1962) Mechanics of thermal contraction cracks and ice wedge polygons in permafrost. *Geol. Soc. Am. Special Paper* , 70. 69 pp.

Lewis, D. W. (1973) Polyphase limestone dikes in the Omaru region, New Zealand. *J. Sedim. Pet.* 43, 1031-1045.

Lewis, W. V. (1954) Pressure release and glacial erosion. *J. Glaciology.* 2, 417-422.

McDonald, B. C. & Shilts, W. W. (1975) Interpretation of faults in glaciofluvial sediments. In: *Glaciofluvial and glaciolacustrine sedimentation.* (Ed. by A. V. Jopling and B. C. McDonald) Spec. Publ. Soc. Econ. Palentol. Miner., Tulsa, 23, 123-131.

Mark, D. M. (1973) Analysis of axial orientation data including till fabrics. *Geol. Soc. Am. Bull.*, 84, 1369-1374.

Peterson, G. L. (1968) Flow structures in sandstone dikes. *Sedim. Geol.* 2, 177-190.

Read, K. O. (1963) The geology of the Princess River area. BSc(hons) thesis. University of Tasmania.

Reimnitz, E. & Marshall, N. F. (1965) Effects of Alaska earthquake and tsunami on recent deltaic sediments. *J. Geophys. Research*, 70, 2363-2376.

Sandford, A. R. (1959) Analytical and experimental study of simple geologic structures. *Geol. Soc. Am. Bull.* 70, 19-52.

Sansom, P. W. (1978) The glacial geomorphology of the Tullah area. BSc(hons) thesis. University of Tasmania. 278 pp.

Shaw, J. (1972) Sedimentation in the ice contact environment with examples from Shropshire, England. *Sedimentology*, 23-62.

Stewart, C. A. (1911) Note on a conglomerate dike in Arizona. *Science*, 33, 431-435.

Thorson, R. M. , Clayton, W.S. & Seeber, L. (1986) Geological evidence for a large prehistoric earthquake in eastern Connecticut. *Geology* 14, 463-467.

Worsley, P. (1975) Till dike genesis: A discussion. *Can. J. Earth Sci.* 12, 1249-1250.